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SPECIAL PUBLICATION

OCEANOGRAPHY FOR LONG RANGE SONAR SYSTEMS

PART I INTRODUCTION TO OCEANOGRAPHY AND PHYSICS OF UNDERWATER SOUND IN THE SEA

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FOREWORD

The success or failure of modern naval operations depends, in part, upon the knowledge and background of the personnel who operate various pieces of hardware in an oceanographic environment, and upon the ability of the operator to understand the interaction between instrument and medium. Successful naval operations require personnel at all levels to have an understanding of the environmental effects on the operation of modern long range sonars and associated weapons systems. This publication provides the necessary background in oceanography and acoustics. Users are encouraged to submit criticism of this publication in the interest of improving future editions.

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INTRODUCTION TO OCEANOGRAPHY

This report is designed to fill a gap between pure and applied oceanography. It is hoped that by bringing the scientist and the military user closer together that understanding of mutual problems will result in solutions that mark important strides forward in the field of military oceanography. The reader must keep in mind that all answers concerning the effects of the environment are not known, and that as knowledge progresses, revisions or additions to this publication will be made.

Annex A, at the end of this report, contains conversion factors and formulas for interchange of units to a form more familiar to each individual reader.

Oceanography is the study of the sea in all its aspects. It includes a study of the physics, chemistry, biology, and geology of the ocean and the interactions at the land-sea and air-sea interfaces. The major properties of the oceans are discussed separately in the pages that follow.

TEMPERATURE

Temperature is a physical property of the oceans and is a measure of the kinetic energy (energy of motion) of molecules. For sea water it may also include the measurement of potential energy due to compression or expansion of the liquid. Techniques for measuring temperatures vary markedly depending upon the use to which the data will be put. Temperatures at depth are measured by reversing thermometers, bathythermographs (BT), or thermistor chains. The system used depends on the accuracy desired, recording procedures, and required speed of acquisition of data.

Inasmuch as the temperature of the atmosphere and the ocean does not change radically from year to year, the heat received from the sun (insolation) must balance, in the same period, the amount lost by reflection and radiation into space. At any one location the amount of insolation depends on such factors as the altitude of the sun, amount of heat absorbed in the atmosphere, and the cloudiness of the region. On a completely overcast, dark, and rainy day the incoming radiation from the sun may be reduced to less than 10% of that received on a clear day. When the sky is partly overcast, however, the amount of radiation received may exceed that on a clear day because of the reflection of heat back to earth from the clouds. In clear oceanic water, about 62% of the incoming energy that reaches the water surface is absorbed within the first meter of water and about 83% is absorbed within 10 meters, whereas in turbid coastal water over 99% is absorbed within the first 10 meters. Thus, in the absence of

other oceanographic and meteorological processes, direct radiation heating would be confined to this 10-meter layer with little or no heat reaching the deeper water.

Heat exchange with the deeper water is caused by stirring of the surface waters by wind. This wind stirring produces a mixed layer of water having the same temperature (isothermal), the thickness of which depends on the force and duration of the wind. At best, however, this wind-induced vertical mixing of surface water reaches a maximum depth of about 300 to 450 meters in winter and about 150 meters in summer.

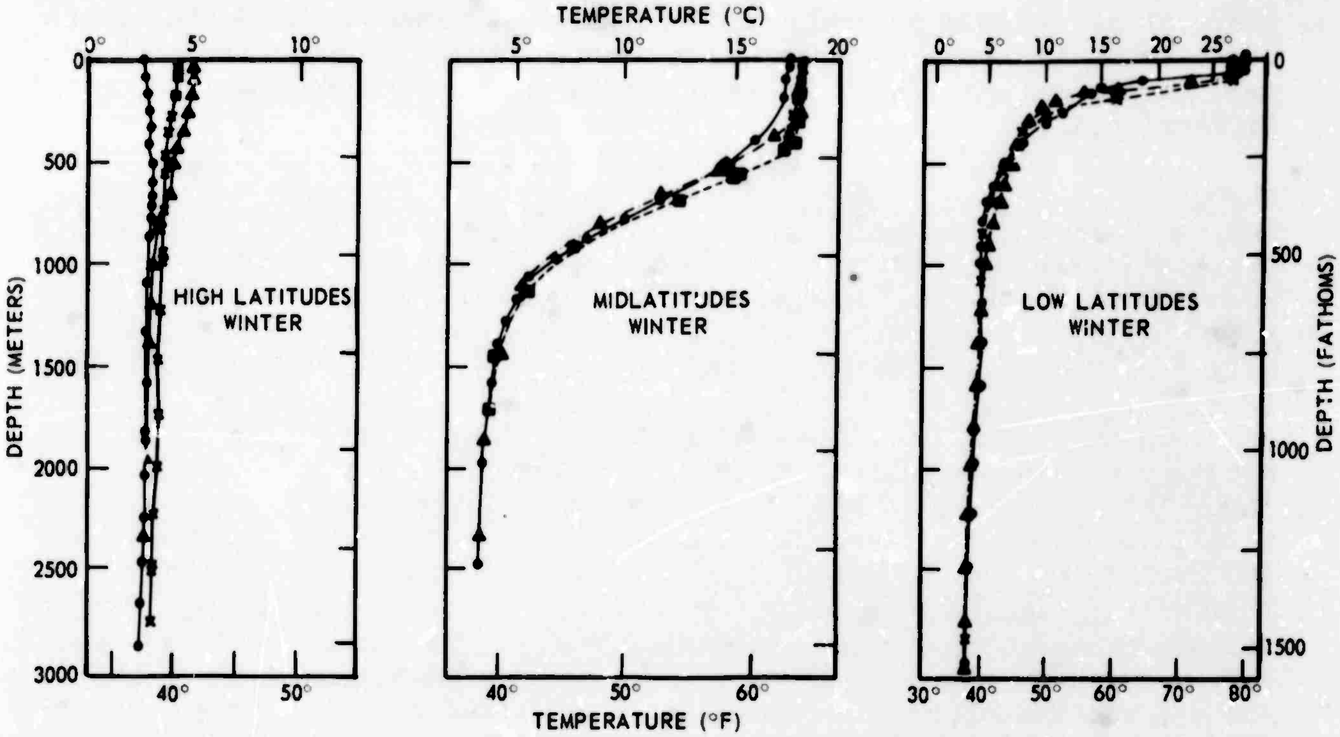
Currents also affect the distribution of temperature. For example, not only does the axis of the Gulf Stream shift with time but masses of the Gulf Stream break off from the edges of the main current and move laterally. These masses may modify temporarily the thermal structure of the waters adjacent to the Gulf Stream. Surface temperatures are warmer on the western sides of the North Atlantic and North Pacific basins than at a corresponding latitude on the eastern sides of these basins. This situation is a result of the clockwise circulation pattern in these ocean areas.

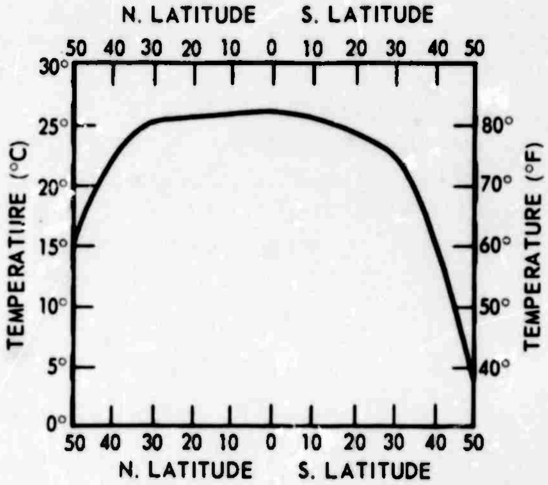
Evaporation and upwelling reduce surface temperatures. Both processes are dependent on winds. In the upwelling process, warmer surface water is blown away from a coast by winds and is replaced by colder subsurface water. Upwelling is especially common along the coasts of Morrocco, southwest Africa, California, and Peru.

As a general rule, the smallest seasonal variations of surface temperature occur in the equatorial and polar regions. Largest variations occur in the midlatitudes. In central oceanic regions, surface temperatures generally differ by about ± 2°C from the monthly mean temperature, although in the vicinity of Newfoundland, and in other regions where warm and cold currents meet, surface temperatures may differ by ± 3 to ± 4°C from the monthly mean.

With few exceptions, water temperature decreases with depth. In high (polar) latitudes, the vertical temperature structure in winter shows relatively little change from surface to bottom (fig. 1). However, in some arctic regions, the surface water in winter may be colder than the subsurface water so that a slight positive gradient (temperature increase with depth) occurs. In summer, lighter, fresher melt water spreads out over the colder water below, and during periods of prolonged calms this layer absorbs and retains the incoming radiation so that its temperature increases rapidly. When the wind increases, this layer of warm water is eventually destroyed by mixing.

In midlatitudes during winter, the water is isothermal to depths of 300 to 500 meters (fig. 1). Below this relatively warm (about 17 to 19°C) isothermal layer, temperatures decrease by about 13°C within a vertical distance of about 800 meters, and then decrease at a fairly uniform rate of about 1°C per 1,000 meters. That part of the vertical structure within which the temperature decreases rapidly with depth is known as the main or permanent thermocline. As the amount of insolation increases, the upper portion of the isothermal layer becomes warmer, and a shallow secondary thermocline forms. This secondary thermocline, known as the seasonal





MEAN SURFACE TEMPERATURE ALONG 30°W LONGITUDE IN THE ATLANTIC OCEAN

FIGURE 1. MEAN SURFACE TEMPERATURE AND REPRESENTATIVE WINTER TEMPERATURE PROFILES

thermocline, usually is shallowest and strongest in summer. As the force of the wind increases in autumn, the seasonal thermocline is deepened by wind mixing and eventually is destroyed as the winter thermal structure is developed.

In low (tropical) latitudes during winter, the vertical temperature structure is similar to that described above, with two major exceptions. One, the mixed isothermal layer is only about 50 to 75 meters thick and, two, the total temperature decrease through the permanent thermocline is about 8°C greater than in midlatitudes. Because of the small seasonal range in surface temperature in low latitudes, summer observations generally show the same vertical thermal distribution as winter observations (fig. 1).

In addition to the seasonal variations that occur within the uppermost layer, diurnal (daily) variations of temperature also occur. In lower latitudes, the diurnal variation can be represented by a sine curve with lowest values between 0200 and 0300 hours and highest values between 1400 and 1500 hours (local time). Measurements in the tropics have shown that with a clear sky and calm, or very light breezes, the average diurnal variation is 1.5°C, whereas the maximum and minimum variation is 1.9 and 1.2°C, respectively. The diurnal increase of surface temperature (and the attendant shallow secondary mixed layer) that occurs between 1400 and 1500 hours sometimes is referred to as the "afternoon effect."

Temperature in the ocean varies widely, both horizontally and with depth. Maximum values of about 32°C are encountered in the Persian Gulf in summer, and the lowest possible values of about -2.0°C (the usual freezing point of sea water) occur in polar regions.

The vertical distribution of temperature in the sea nearly everywhere shows a decrease of temperature with depth. Since colder water is denser, it sinks below warmer water. This results in a temperature distribution just opposite to that in the earth's crust, where temperature increases with depth below the surface of the ground. Envelopes of the vertical distribution of temperature for broad ocean areas are shown in Figure 2.

SALINITY

Salinity is a physical property of the oceans and is a measure of all the dissolved material in sea water. Salinity is a weight to weight relationship (grams per kilogram) and is expressed as the number of parts of dissolved material in a thousand parts of sea water (‰).

The ratios of the major dissolved constituents in sea water are practically constant (law of constancy of composition); therefore, an analysis of any one easily measured ion (chlorine) can be used as a measure of the total salt content of a sample. As the direct measurement of salinity is difficult and tedious, oceanographers have adopted the easier and sufficiently accurate method of measuring the chlorinity (1.00045 times the chlorine content) and computing the salinity from the empirical formula.

$$\text{Salinity (‰)} = 0.03 + 1.805 \times \text{Chlorinity (‰)}$$

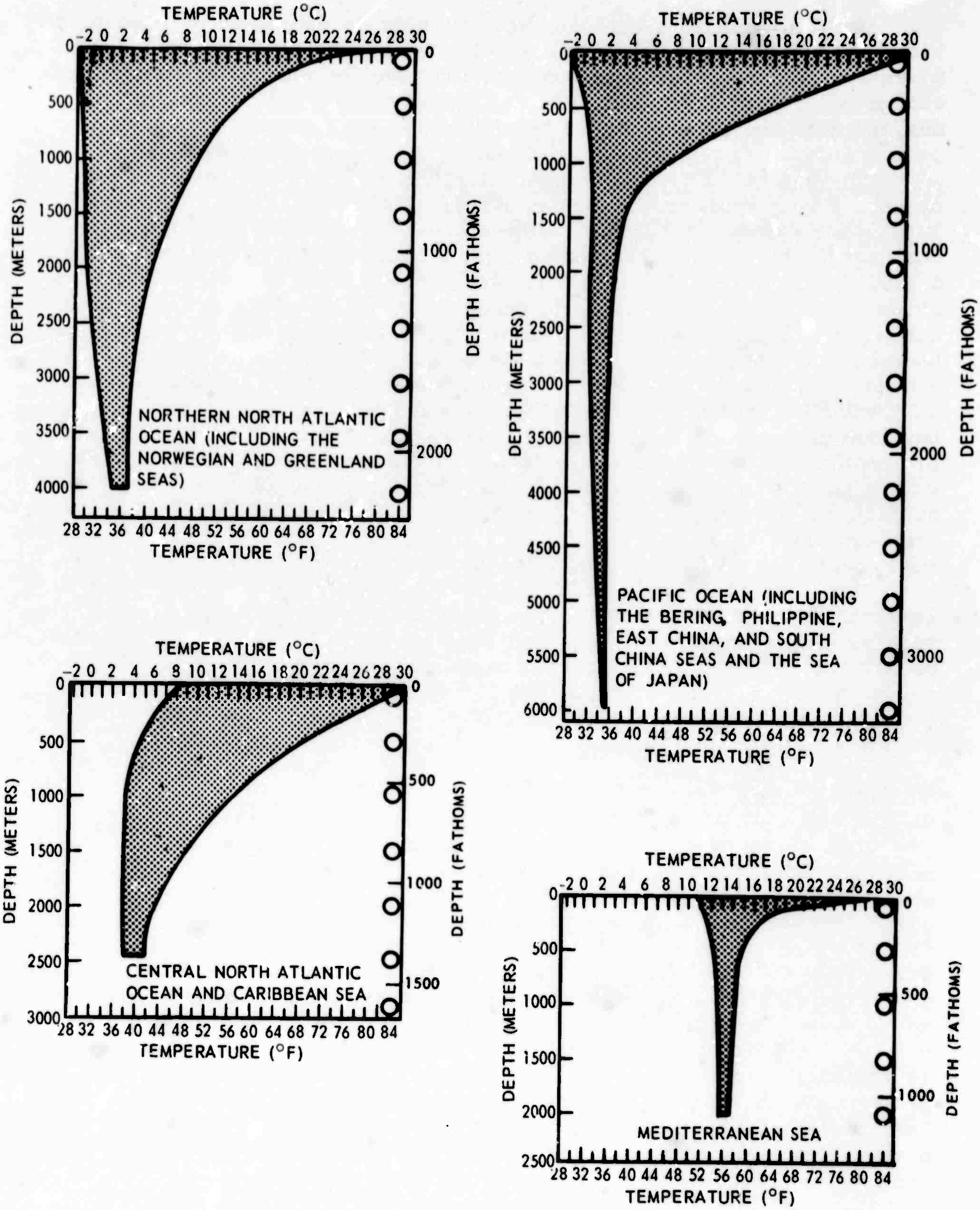


FIGURE 2. TEMPERATURE RANGES WITH DEPTH

In open oceanic areas, surface salinity is decreased by precipitation, increased by evaporation, and changed by vertical and horizontal advection. In nearshore regions, salinity may be reduced by river discharge and runoff from land. In regions where ice occurs, salinities may increase during periods of ice formation and decrease during ice melt.

Surface salinity varies with latitude in a similar manner in all oceans. Maximum salinities occur between 20° and 23° north and south latitudes, whereas minimum salinities occur near the Equator and toward the higher latitudes. The controlling factor in average surface salinity distribution is the difference between evaporation and precipitation as shown in Figure 3; the effect of ocean currents is of minor importance. Exceptions to this general statement do occur and local variations should be expected, as for example near the mouths of such rivers as the Amazon, Rio de la Plata, and the Mississippi, and in the Atlantic coastal waters of the United States, Labrador, Spain, and Scandinavia. Probably the best known region of strong horizontal salinity gradients is the region of the Grand Banks where the warm, saline Gulf Stream waters mix with the waters of the colder, less saline Labrador Current. Here, water with a salinity as low as 32‰ may override or lie adjacent to water of salinity greater than 36‰. A similar situation exists in the Pacific Ocean in the region northeast of Japan where the Kuroshio and the Oyashio Currents mix.

Positive (increasing with depth) salinity gradients tend to develop in latitudes higher than 40° north and south where, generally, precipitation exceeds evaporation. In summer, these positive salinity gradients usually are accompanied by strong negative temperature gradients. Vertical stability (strong positive density gradient) of the water column (especially in coastal regions) results, and these strong shallow salinity (and temperature) gradients persist throughout the summer.

SOUND SPEED

The speed of sound in the ocean is usually computed from measurements of temperature, salinity, and depth (pressure); however, new devices are being developed to measure sound speed directly. Sound speed in the ocean increases with an increase of temperature, salinity, or pressure. Of the three, temperature has the greatest influence on sound speed. With a typical distribution of sound speed with depth (Fig. 4), the speed decreases with depth to a minimum, below which the speed increases (primarily as a function of pressure since temperature and salinity below the minimum level do not change radically). The point of minimum sound speed is called the sound channel axis. Sound trapped in this channel is propagated long distances. The depth of the channel axis varies with latitude as shown in Figure 5. The nomogram presented in Figure 6 permits a rapid computation of sound speed when temperature, salinity, and depth are known.

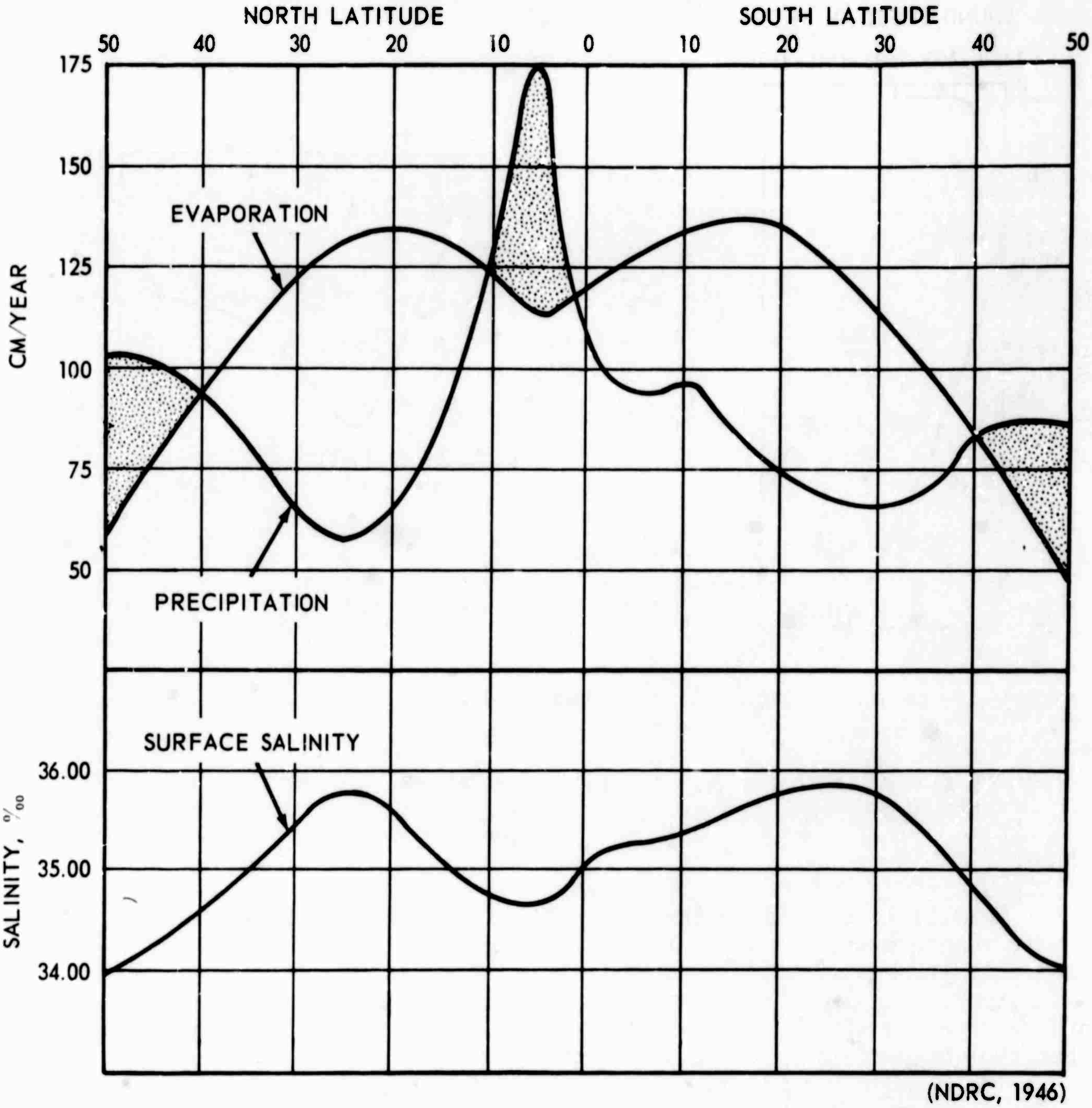
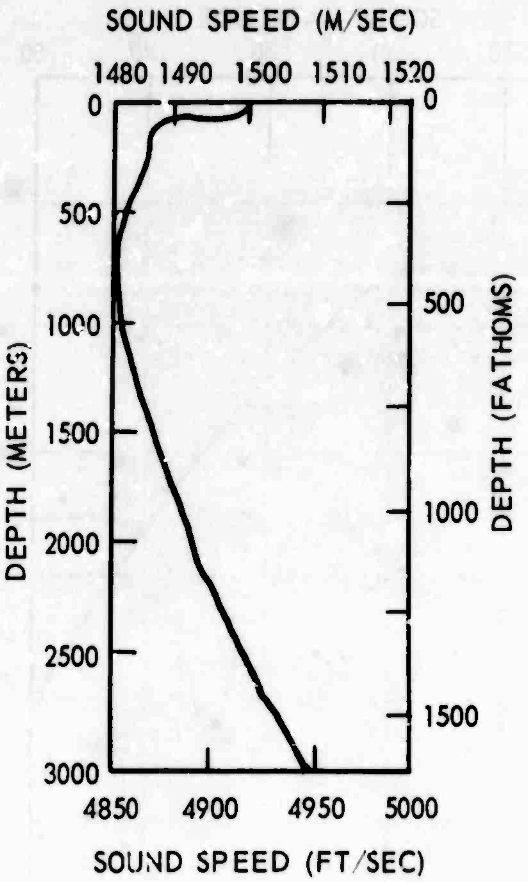


FIGURE 3. VARIATION OF AVERAGE EVAPORATION, PRECIPITATION, AND SALINITY WITH LATITUDE. SHADED AREAS SHOW REGIONS WHERE PRECIPITATION EXCEEDS EVAPORATION.



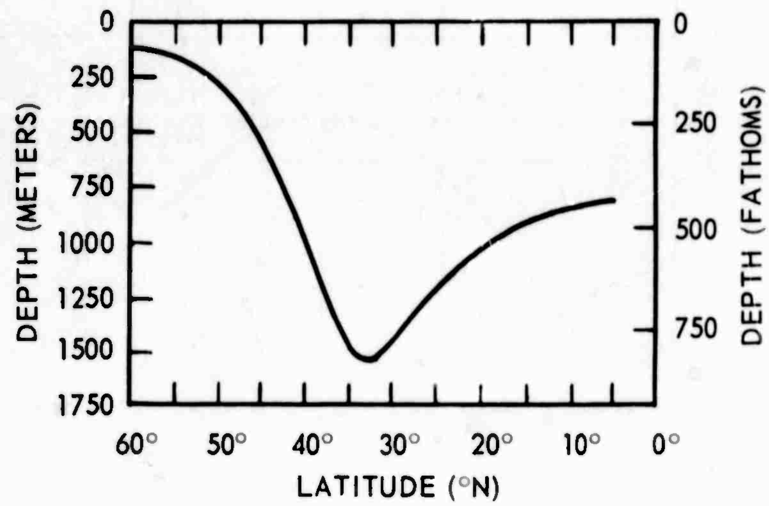


FIGURE 4. TYPICAL PROFILE OF SOUND SPEED

FIGURE 5. VARIATION IN DEPTH OF SOUND CHANNEL AXIS WITH LATITUDE IN THE NORTH ATLANTIC OCEAN

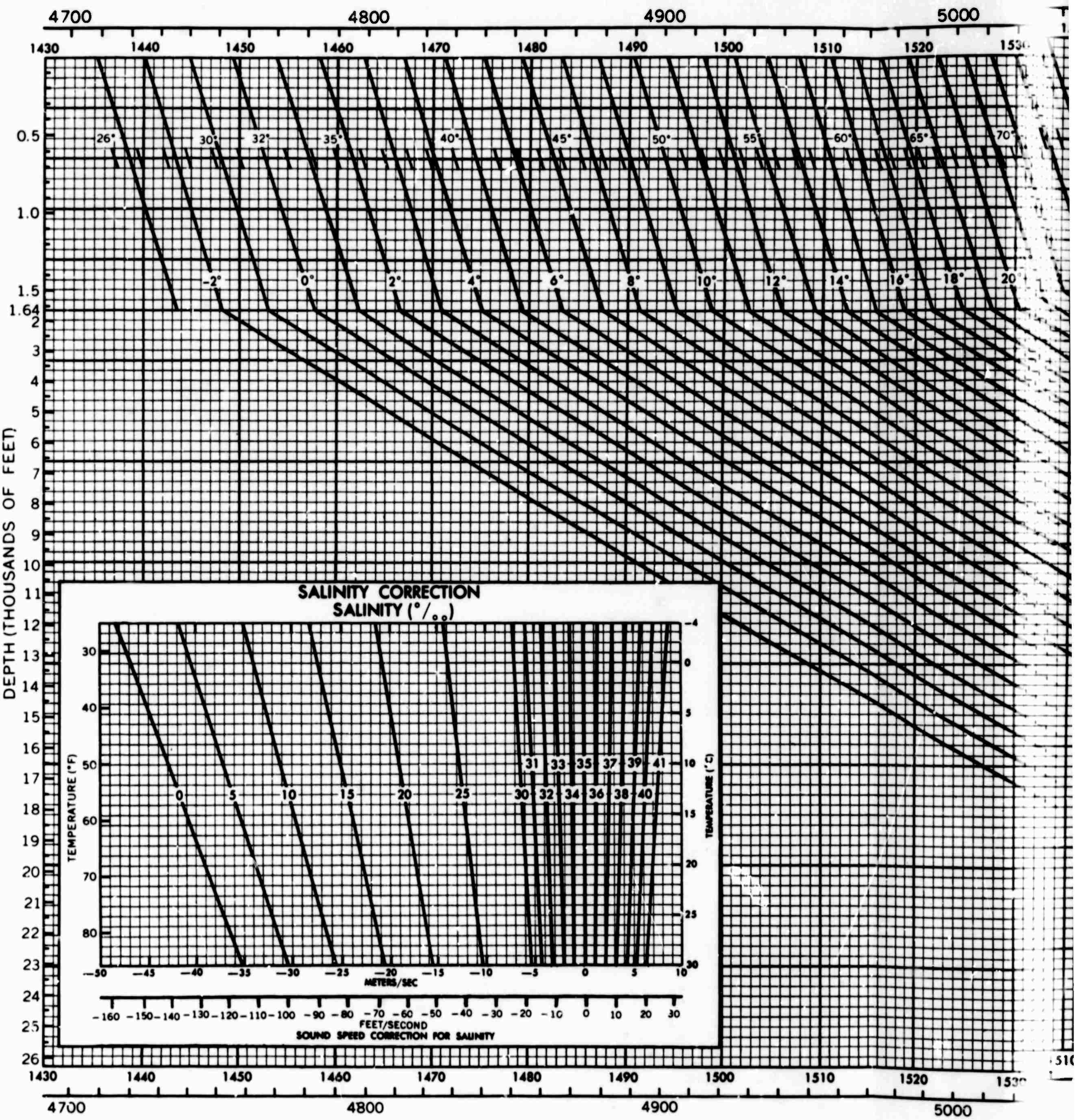
DENSITY

Density is defined as the ratio of the mass of a portion of matter to its volume (mass/volume). The units of density depend upon the units used to express mass and volume (grams per cubic centimeter).

Density is usually computed from the known temperature, salinity, and pressure at a given point. Density of sea water is slightly greater than that of pure water by virtue of the dissolved substances present in sea water. Conversion from density ρ to a simplified value called sigma-t (σ_t) is accomplished by $\sigma_t = (\rho - 1)\ 10^3$.

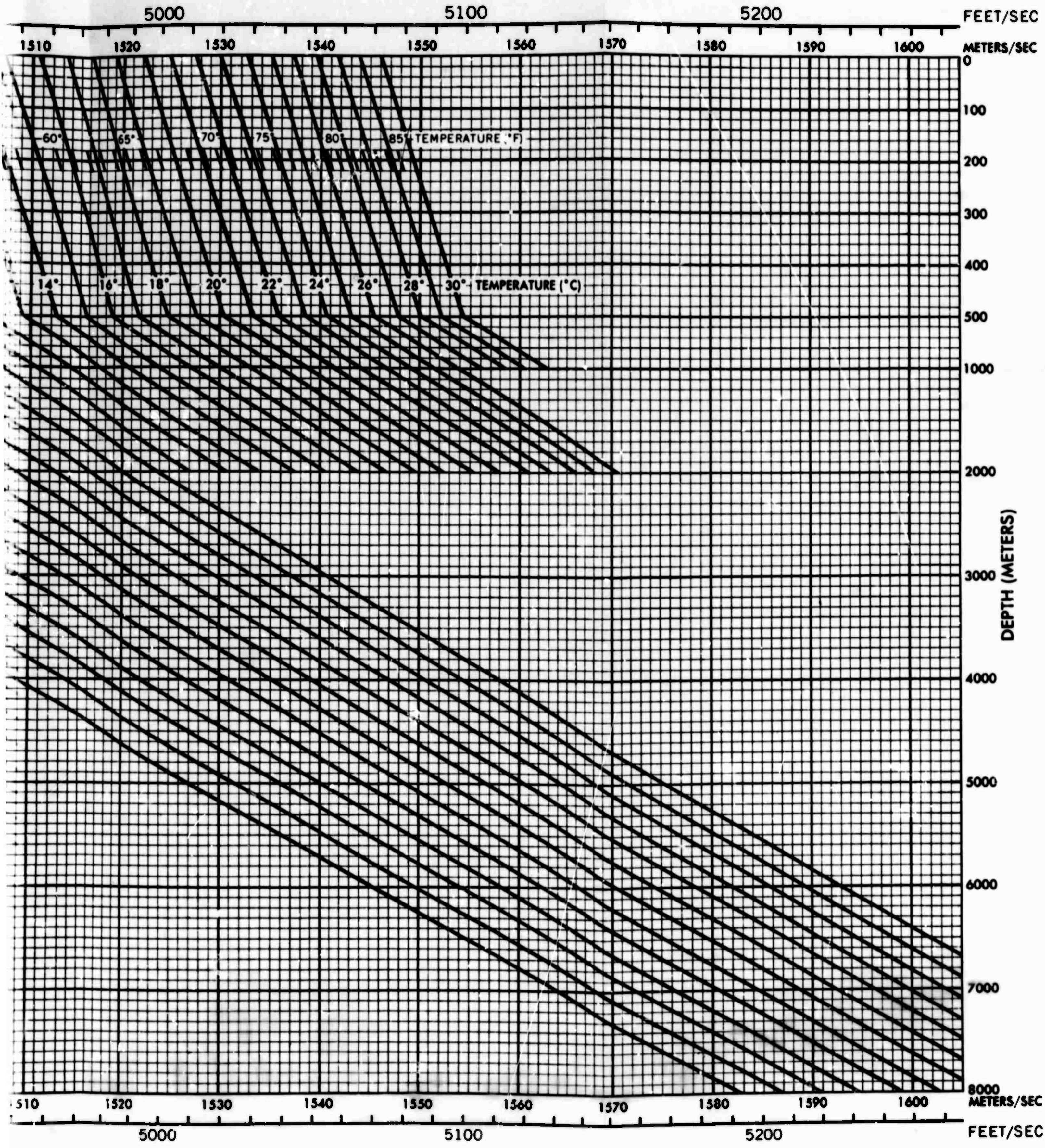
Thus if $\rho = 1.02577$ then $\sigma_t = 25.77$. The expression σ_t has the advantage of containing fewer digits and is a significant value because any water motion along a σ_t surface involves little energy loss.

All processes that affect the distribution of temperature and salinity also affect the distribution of density. The greatest changes in density of sea water occur at the surface, where the water is subject to influences not present at depth. Density is decreased by precipitation, runoff from land, melting of ice, or heating. When the surface



A

FIGURE 6. SOUND SPEED NOMOGRAM (BASED O



ED NOMOGRAM (BASED ON WILSON'S EQUATION)

B

water becomes less dense it tends to float on top of the more dense water below (stable condition). The density of surface water is increased by evaporation, formation of sea ice, and by cooling. If the surface water becomes more dense than that below, it sinks to the level of waters of the same density. Here it tends to spread out to form a layer, or to increase the thickness of the layer. Less dense water rises to make room, and the surface water moves in to replace that which has descended. Thus, a convective circulation forms to re-establish equilibrium. If the surface water becomes sufficiently dense, it sinks all the way to the bottom. If this occurs in an area where horizontal flow is unobstructed, the water which has descended spreads to other regions, creating a dense bottom layer. Since the greatest increase in density occurs in polar regions, where the air is cold and great quantities of ice form, the cold, dense polar water sinks to the bottom by vertical convection currents and then spreads to lower latitudes.

In offshore equatorial and temperate zones, the temperature of the surface water is so high that the density of the water remains low even in regions where excess evaporation causes high salinities. As a result, convection currents are limited to a relatively thin layer near the surface and the stability of the remainder of the water column is unaffected.

Layers of homogeneous water are formed in regions where strong winds occur, such as the trade wind belts. In such regions the wind is capable of overcoming the density gradient and causes mechanical wind mixing of the surface waters. The depth to which wind mixing penetrates increases with increasing wind force.

PRESSURE

Pressure is generally expressed in units of the centimeter-gram-second system. The basic unit of this system is one dyne per square centimeter. This is a very small unit, one million constituting a practical unit called a bar, which is nearly equal to one atmosphere. Atmospheric pressure is often expressed in terms of millibars, 1,000 of these being equal to 1 bar. In oceanographic work, water pressure is commonly expressed in terms of decibars, 10 of these being equal to 1 bar. One decibar is equal to nearly 1 1/2 pounds per square inch. The decibar unit is convenient because it is very nearly the pressure exerted by 1 meter of water. Thus, the pressure in decibars is approximately the same as the depth in meters. The increase in pressure with depth is nearly constant because water is only slightly compressible.

COMPRESSIBILITY

Compressibility is the relative decrease of the volume of a system with increasing pressure assuming isothermal conditions. Sea water is nearly incompressible, its coefficient of compressibility being only 0.000046 per bar under standard conditions. This value changes slightly with changes of temperature or salinity. The effect of compression is to

force the molecules of the substance closer together, causing an increase in density. Even though the compressibility of sea water is low, its total effect is considerable because of the large amount of water involved. If the compressibility of sea water were zero, sea level would be about 30 meters higher than it now is.

VISCOSITY

Viscosity is resistance to flow or deformation. Viscosity increases with increasing salinity, but the effect is not as marked as the increase that occurs with decreasing temperature. The rate is not uniform, becoming greater as the temperature decreases.

SEA AND SWELL

Waves on the surface of the sea are caused principally by wind, but other factors, such as submarine earthquakes, volcanic eruptions, and the tide also cause waves. If a breeze of less than 2 knots starts to blow across smooth water, small wavelets called ripples form almost instantaneously. When the breeze dies, the ripples disappear as suddenly as they formed, the level surface being restored by surface tension of the water. If the wind speed exceeds 2 knots, more stable waves (gravity waves) gradually form and progress with the wind.

While the generating wind blows, the resulting waves are referred to as sea. When the wind stops or changes direction, the waves that travel on without relation to local winds are called swell. Table 1 gives some of the relationships involved.

Unlike wind and current, waves are not deflected appreciably by the rotation of the earth, but move in the direction in which the generating wind blows. When this wind ceases, friction and spreading cause the waves to be reduced in height, or attenuated, as they move across the surface. However, the reduction takes place so slowly that a swell continues until it reaches some obstruction, such as a shore. When sufficient data on wind conditions are available, the swell and state of the sea can be predicted in advance.

Ocean waves are very nearly the shape of an inverted cycloid, the figure formed by a point inside the rim of a wheel rolling along a level surface. This shape is shown in Figure 7. The highest parts of waves are called crests, and the intervening lowest parts, troughs. Since the crests are steeper and narrower than the troughs, the mean or still water level is a little lower than halfway between the crests and troughs. The vertical distance between trough and crest is called wave height (H). The horizontal distance between successive crests, measured in the direction of travel, is called wave length (L). The time interval between passage of successive crests at a fixed point is called the wave period (P). Wave height, length, and period depend upon a number of factors, such as the wind speed, the length of time it has blown, and its fetch (the straight distance it has traveled over the surface).

Sea State	Sea Description	Wind Force (Beaufort)	Description	Range (Knots)	Wind Velocity (Knots)	Wave Height (Feet): Average	Significant	Average 1/10 Highest	Significant Range of Periods (Seconds)	T max (Period of Maximum Energy of Spectra)	T̄ (Average Period)	L̄ (Average Wave Length)	Minimum Fetch (Nautical Miles)	Minimum Duration (Hours)	E Value
0	SEA LIKE A MIRROR.	0	CALM	<1	0**	0	0	0	—	—	—	—	—	—	—
	RIPPLES WITH THE APPEARANCE OF SCALES ARE FORMED, BUT WITHOUT FOAM CRESTS.	1	LIGHT AIRS	1-3	2	0.05	0.08	0.10	TO 1.2	0.7	0.5	10 IN	5	18 MIN	0.00
1	SMALL WAVELETS, STILL SHORT BUT MORE PRONOUNCED; CRESTS HAVE A GLASSY APPEARANCE, BUT DO NOT BREAK.	2	LIGHT BREEZE	4-6	5	0.18	0.29	0.37	0.4-2.8	2.0	1.4	6.7 FT	8	39 MIN	0.02
	LARGE WAVELETS. CRESTS BEGIN TO BREAK. FOAM OF GLASSY APPEARANCE. PERHAPS SCATTERED WHITE HORSES.	3	GENTLE BREEZE	7-10	8.5	0.6	1.0	1.2	0.8-5.0	3.4	2.4	20	9.8	1.7 HRS	0.09
					10	0.88	1.4	1.8	1.0-6.0	4.0	2.9	27	10	2.4	0.22
2	SMALL WAVES, BECOMING LARGER; FAIRLY FREQUENT WHITE HORSES.	4	MODERATE BREEZE	11-16	12	1.4	2.2	2.8	1.0-7.0	4.8	3.4	40	18	3.8	0.33
					13.5	1.8	2.9	3.7	1.4-7.6	5.4	3.9	52	24	4.8	0.65
3					14	2.0	3.3	4.2	1.5-7.8	5.6	4.0	59	28	5.2	1.30
					16	2.9	4.6	5.8	2.0-8.8	6.5	4.6	71	40	6.6	1.40
4	MODERATE WAVES, TAKING A MORE PRONOUNCED LONG FORM; MANY WHITE HORSES ARE FORMED. (CHANCE OF SOME SPRAY).	5	FRESH BREEZE	17-21	18	3.8	6.1	7.8	2.5-10.0	7.2	5.1	90	55	8.3	2.70
					19	4.3	6.9	8.7	2.8-10.6	7.7	5.4	99	65	9.2	4.70
					20	5.0	8.0	10	3.0-11.1	8.1	5.7	111	75	10	6.30
5	LARGE WAVES BEGIN TO FORM; THE WHITE FOAM CRESTS ARE MORE EXTENSIVE EVERYWHERE. (PROBABLY SOME SPRAY).	6	STRONG BREEZE	22-27	22	6.4	10	13	3.4-12.2	8.9	6.3	134	100	12	7.70
					24	7.9	12	16	3.7-13.5	9.7	6.8	160	130	14	10.30
					24.5	8.2	13	17	3.8-13.6	9.9	7.0	164	140	15	16.30
6					26	9.6	15	20	4.0-14.5	10.5	7.4	188	180	17	21.50
	SEA HEAPS UP AND WHITE FOAM FROM BREAKING WAVES BEGINS TO BE BLOWN IN STREAKS ALONG THE DIRECTION OF THE WIND. (SPINDRIFT BEGINS TO BE SEEN).	7	MODERATE GALE	28-33	28	11	18	23	4.5-15.5	11.3	7.9	212	230	20	30.70
					30	14	22	28	4.7-16.7	12.1	8.6	250	280	23	41.00
					30.5	14	23	29	4.8-17.0	12.4	8.7	258	290	24	50.70
					32	16	26	33	5.0-17.5	12.9	9.1	285	340	27	61.00
7	MODERATELY HIGH WAVES OF GREATER LENGTH; EDGES OF CRESTS BREAK INTO SPINDRIFT. THE FOAM IS BLOWN IN WELL MARKED STREAKS ALONG THE DIRECTION OF THE WIND. SPRAY AFFECTS VISIBILITY.	8	FRESH GALE	34-40	34	19	30	38	5.5-18.5	13.6	9.7	322	420	30	81.10
					36	21	35	44	5.8-19.7	14.5	10.3	362	500	34	111.00
					37	23	37	46.7	6-20.5	14.9	10.5	376	530	37	140.00
					38	25	40	50	6.2-20.8	15.4	10.7	392	600	38	160.00
					40	28	45	58	6.5-21.7	16.1	11.4	444	710	42	192.00
8	HIGH WAVES. DENSE STREAKS OF FOAM ALONG THE DIRECTION OF THE WIND. SEA BEGINS TO ROLL. VISIBILITY AFFECTED.	9	STRONG GALE	41-47	42	31	50	64	7-23	17.0	12.0	492	830	47	232.00
					44	36	58	73	7-24.2	17.7	12.5	534	960	52	316.00
					46	40	64	81	7-25	18.6	13.1	590	1110	57	399.00
	VERY HIGH WAVES WITH LONG OVERHANGING CRESTS. THE RESULTING FOAM IS IN GREAT PATCHES AND IS BLOWN IN DENSE WHITE STREAKS ALONG THE DIRECTION OF THE WIND. ON THE WHOLE THE SURFACE OF THE SEA TAKES A WHITE APPEARANCE. THE ROLLING OF THE SEA BECOMES HEAVY AND SHOCKLIKE. VISIBILITY IS AFFECTED.	10	WHOLE GALE*	48-55	48	44	71	90	7.5-26	19.4	13.8	650	1250	63	498.00
					50	49	78	99	7.5-27	20.2	14.3	700	1420	69	616.00
					51.5	52	83	106	8-28.2	20.8	14.7	736	1560	73	718.00
					52	54	87	110	8-28.5	21.0	14.8	750	1610	75	870.00
					54	59	95	121	8-29.5	21.8	15.4	810	1800	81	919.00
	EXCEPTIONALLY HIGH WAVES (SMALL AND MEDIUM-SIZED SHIPS MIGHT FOR A LONG TIME BE LOST TO VIEW BEHIND THE WAVES.) THE SEA IS COMPLETELY COVERED WITH LONG WHITE PATCHES OF FOAM LYING ALONG THE DIRECTION OF THE WIND. EVERYWHERE THE EDGES OF THE WAVE CRESTS ARE BLOWN INTO FROTH. VISIBILITY AFFECTED.	11	STORM*	56-63	56	64	103	130	8.5-31	22.6	16.3	910	2100	88	1110.00
					59.5	73	116	148	10-32	24.0	17.0	985	2500	101	1331.00
	AIR FILLED WITH FOAM AND SPRAY. SEA COMPLETELY WHITE WITH DRIVING SPRAY; VISIBILITY VERY SERIOUSLY AFFECTED.	12	HURRICANE*	64-71	>64	>80*†	>128*†	>164*†	10-(35)	(26)	(18)	—	—	—	—

* FOR HURRICANE WINDS (AND OFTEN WHOLE GALE AND STORM WINDS) REQUIRED DURATIONS AND FETCHES ARE RARELY ATTAINED. SEAS ARE THEREFORE NOT FULLY ARISEN.

** A HEAVY BOX AROUND THIS VALUE MEANS THAT THE VALUES TABULATED ARE AT THE CENTER OF THE BEAUFORT RANGE.

† FOR SUCH HIGH WINDS, THE SEAS ARE CONFUSED. THE WAVE CRESTS BLOW OFF, AND THE WATER AND THE AIR MIX.

TABLE 1. WIND AND SEA RELATIONSHIPS

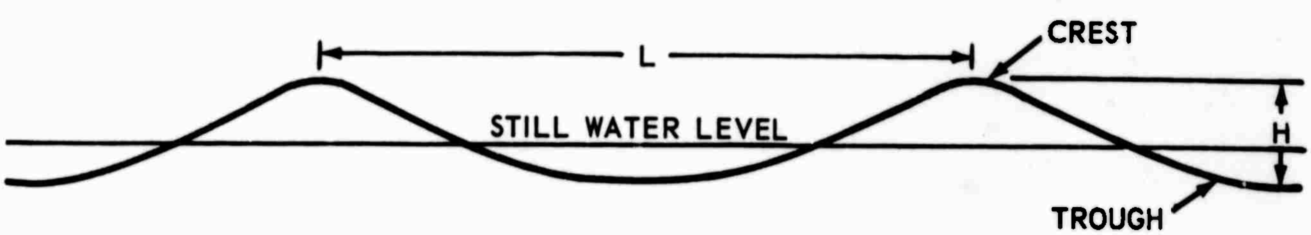


FIGURE 7. TYPICAL SEA WAVE

If the water is deeper than one-half the wave length (L), this length in feet is theoretically related to period (P) in seconds by the

formula

$$L = 5.12 P^2.$$

The actual value has been found to be a little less than this for swell, and about two-thirds the length determined by this formula for sea. When the waves leave the generating area and continue as free waves, the wave length and period continue to increase, while the height decreases.

The speed (S) of a free wave in deep water is nearly independent of its height or steepness. For swell, the relationship in knots to the period (P) in seconds is given by the formula

$$S = 3.03 P.$$

The theoretical relationship between speed, wave length, and period is shown in Figure 8. As waves continue on beyond the generating area the period, length, and speed all increase, providing some indication of the distance from the generating area. The time needed for a wave system to travel a given distance is double that which would be indicated by the speed of individual waves. This is because the front wave gradually disappears and transfers its energy to succeeding waves. The process is followed by each front wave in succession at such a rate that the wave system advances at a speed (group velocity) which is just half that of the individual waves.

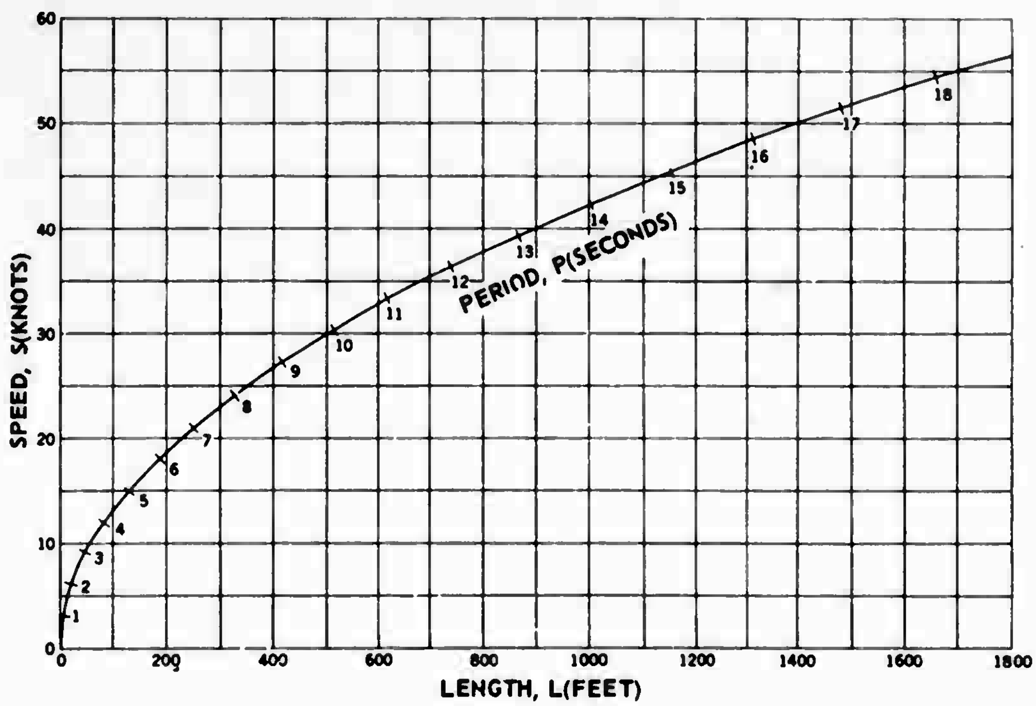
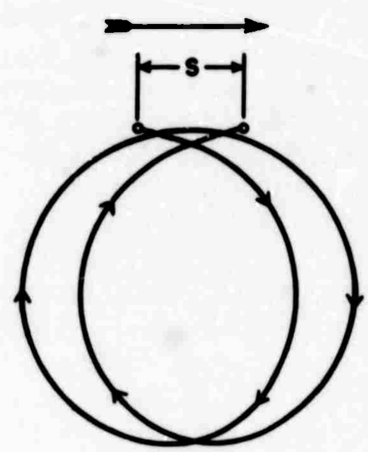


FIGURE 8. RELATIONSHIP BETWEEN SPEED, LENGTH, AND PERIOD OF WAVES IN DEEP WATER

Because of the existence of many independent wave systems at the same time, the sea surface acquires a complex and irregular pattern. Also, since the longer waves (longer periods and faster speeds) outrun the shorter ones (shorter periods and slower speeds), the resulting interference adds to the complexity of the pattern and is the principle reason that successive waves are not of the same height. The irregularity of the surface may be further accentuated by the presence of wave systems crossing at an angle to each other.

In reporting average wave heights, the mariner has a tendency to neglect the lower ones. It has been found empirically that the reported value is about the average of the one-third highest waves. This is sometimes called the "significant" wave height. The approximate relationship between this height and others, is as follows:

Wave	Relative Height
Average	0.64
Significant	1.00
Highest 10%	1.29
Highest	1.87



ORBITAL MOTION AND DISPLACEMENT, S, OF A PARTICLE ON THE SURFACE OF DEEP WATER DURING TWO WAVE PERIODS.

FIGURE 9. ORBITAL MOTION

Figure 9 shows that the path of a particle of water on the surface of the ocean follows a somewhat circular orbit as a wave passes, but moves very little in the direction of motion of the wave. As the crest passes,

the particle moves forward, giving the water the appearance of moving with the wave. As the trough passes, the motion is in the opposite direction. The radius of each circular orbit decreases with depth, approaching zero at a depth equal to about half the wave length. In shallower water the orbits become more elliptical, and in very shallow water, as at a beach, the vertical motion disappears almost completely.

Since the speed is greater at the top of the orbit than at the bottom the particle is not exactly at its original point following passage of a wave, but has moved slightly in the direction of motion of the wave. However, since this advance is small in relation to the vertical displacement, a floating object is raised and lowered by passage of a wave but moved little from its original position.

A following current affects waves by increasing wave lengths and decreasing wave heights. An opposing current has the opposite effect, decreasing the length and increasing the height. A strong opposing current may cause the waves to break. The extent of wave alteration is dependent upon the ratio of the still-water wave speed to the speed of the current. Moderate ocean currents running at oblique angles to wave direction appear to have little effect, but strong tidal currents perpendicular to a system of waves have been observed to completely destroy them in a short period of time.

The potential energy of a wave is related to the vertical distance of each particle from its still-water position, and this energy, therefore, moves with the wave. In contrast, the kinetic energy of a wave is related to the speed of the particles, being distributed evenly along the entire wave. The amount of kinetic energy in even a moderate wave is tremendous. A 4-foot, 10-second wave striking a coast expends more than 35,000 horsepower per mile of beach. An increase in temperature of the water in the relatively narrow surf zone in which this energy is expended would seem to be indicated, but no pronounced increase has been measured. Apparently, any heat that may be generated is dissipated to the deeper water beyond the surf zone.

When ice crystals form in sea water, internal friction is greatly increased. This results in smoothing of the sea surface. The effect of pack ice is even more pronounced. A vessel following a lead through such ice may be in smooth water even when a gale is blowing and heavy seas are beating against the outer edge of the pack. Hail as well as sea slicks are also effective in flattening the sea, even in a high wind.

Tsunamis are ocean waves produced by a sudden, large-scale motion of a portion of the ocean floor or the shore, as a volcanic eruption, earthquake (sometimes called a seaquake if it occurs at sea), or landslide. Either a tsunami or a storm wave that overflows the land is popularly called a tidal wave, although it bears no relation to the tide.

If a volcanic eruption occurs below the surface of the sea, the escaping gases cause a quantity of water to be pushed upward in the shape of a dome or mound. The same effect is caused by the sudden raising of a portion of the bottom. As this water settles back, it creates a wave (tsunami) which travels at high speed across the surface of the ocean. Tsunamis usually occur in series, gradually increasing in height until a

maximum is reached between about the third and eighth wave. Waves may continue to form for several hours, or even for days.

In deep water the wave height of a tsunami is probably never greater than 2 or 3 feet. Since the wave length is usually considerably more than 100 miles, the wave is not conspicuous at sea. In the Pacific, where most tsunamis occur, the wave period varies between about 15 and 60 minutes, and the speed in deep water is more than 400 knots.

Earthquakes below the surface of the sea may produce a longitudinal wave that travels upward toward the surface at the speed of sound in the medium. When a ship encounters such a wave, it is felt as a sudden shock which may be of such severity that the crew thinks the vessel has struck bottom.

Thus far, the discussion has been confined to waves on the surface of the sea. Internal waves, or boundary waves, are created below the surface at the boundaries between water strata of different densities. The density differences between adjacent water strata in the sea are considerably less than that between sea and air; consequently, internal waves are much more easily formed than surface waves, and they are often much larger. The maximum height of wind waves on the surface is about 60 feet, but internal wave heights as great as 300 feet may occur.

Internal waves are usually detected by time-series observations of the vertical temperature distribution. Internal waves have periods as short as a few minutes, and as long as 4 or 5 days. The full significance of internal waves has not been determined, but it is known that they may cause submarines to rise and fall like a ship at the surface, and they also seriously affect sound transmission in the sea.

CURRENTS AND MASS TRANSPORT

Oceanic circulation plays a major role in the distribution of many environmental factors that affect military oceanography. Although the general directions and speeds of the principal permanent surface currents are fairly well known, knowledge of surface currents at any given time and place is lacking. Likewise, unfortunately, knowledge of subsurface currents is seriously limited. Recent important developments in current measuring devices and new current investigating techniques have led to some significant discoveries in subsurface circulation. For example, a southward flow has been found at great depths beneath and lateral to the powerful north-setting Gulf Stream, and an unexplained jet of eastward-flowing subsurface water (the Cromwell Current) with speeds to 3 knots has been discovered setting counter to the west-setting North Equatorial Current in the Pacific Ocean. However, direct subsurface current measurements in the ocean are not nearly abundant enough to establish a pattern of subsurface circulation. Our knowledge of subsurface circulation is based primarily on a method of water density computations from the distribution of temperature and salinity, factors much easier to measure and for which much more abundant data are available. Current charts that are constructed from these computations are generalized, but knowledge of subsurface currents would be even more scanty if based entirely on direct observations.

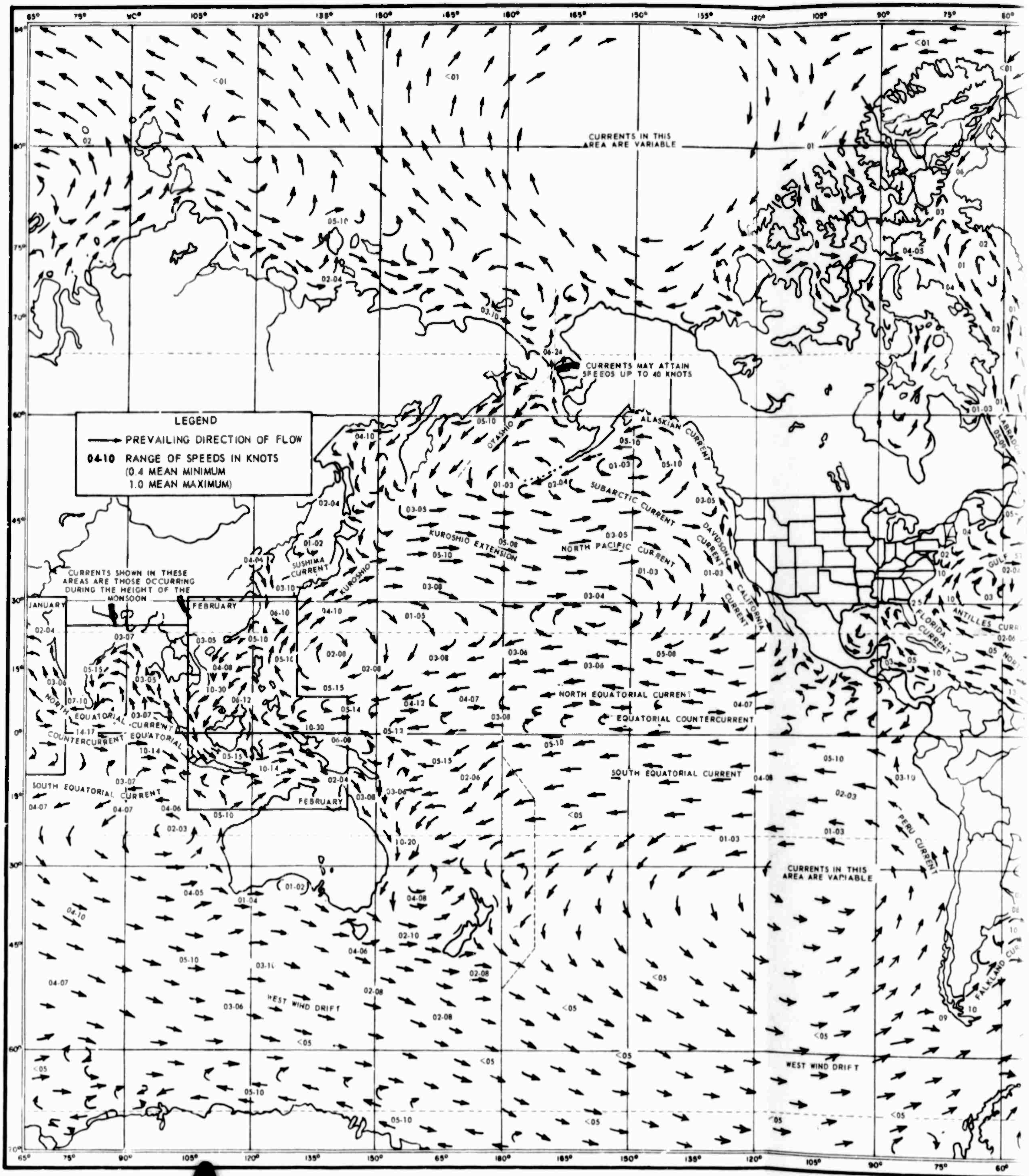
Currents in the sea generally are produced by wind, tide, differences in density between water masses, sea level differences, or runoff from the land. Currents may be classified roughly as wind driven, tidal, geopotential, or hydraulic, but are usually some combination of these. Wind-driven currents are those initiated and sustained by the force of the wind exerting stress on the sea surface. This stress causes the surface water to move, and this movement is transmitted to the underlying water to a depth which is dependent mainly on the strength and persistence of the wind and internal friction. Tidal currents are the horizontal expression of the tidal forces and are especially significant in shallow water, where they often become strong or the predominant flow. In deep water, tidal currents could be the primary component in some areas, and the effects may be significant at great depths. Geopotential currents are associated with density differences in water masses. These differences may be used to determine currents to great depths in the oceans. Hydraulic currents are caused by differences in sea level between two water bodies. These currents usually occur in straits separating water bodies.

Currents normally are measured by their speed in knots and by direction according to compass points or degrees. Observations of currents are made directly by mechanical devices that record speed and direction or indirectly by water density computation, drift bottles, or visually (slicks and water color differences).

The systems of currents an the oceans of the world keeps the water continually in circulation. The midwinter and midsummer positions of these major ocean currents (general circulation) are shown in Figures 10 and 11. These positions shift only slightly with seasons, except in the northern part of the Indian Ocean and along the China coast where the monsoons cause the currents to flow in opposite directions in winter and summer. Currents appear on most charts as well behaved continuous streams defined by clear boundaries and with gradually changing directions. These representations usually are smoothed patterns derived from averages of many observations and do not represent synoptic or instantaneous situations. Figure 12 is an example, based on actual observations of the meandering of the Gulf Stream that can be expected within short periods of time. Unfortunately, prediction of such meanders cannot be made at present. Thus, surface and subsurface current charts must represent resultant or prevailing currents derived statistically and based on many observations.

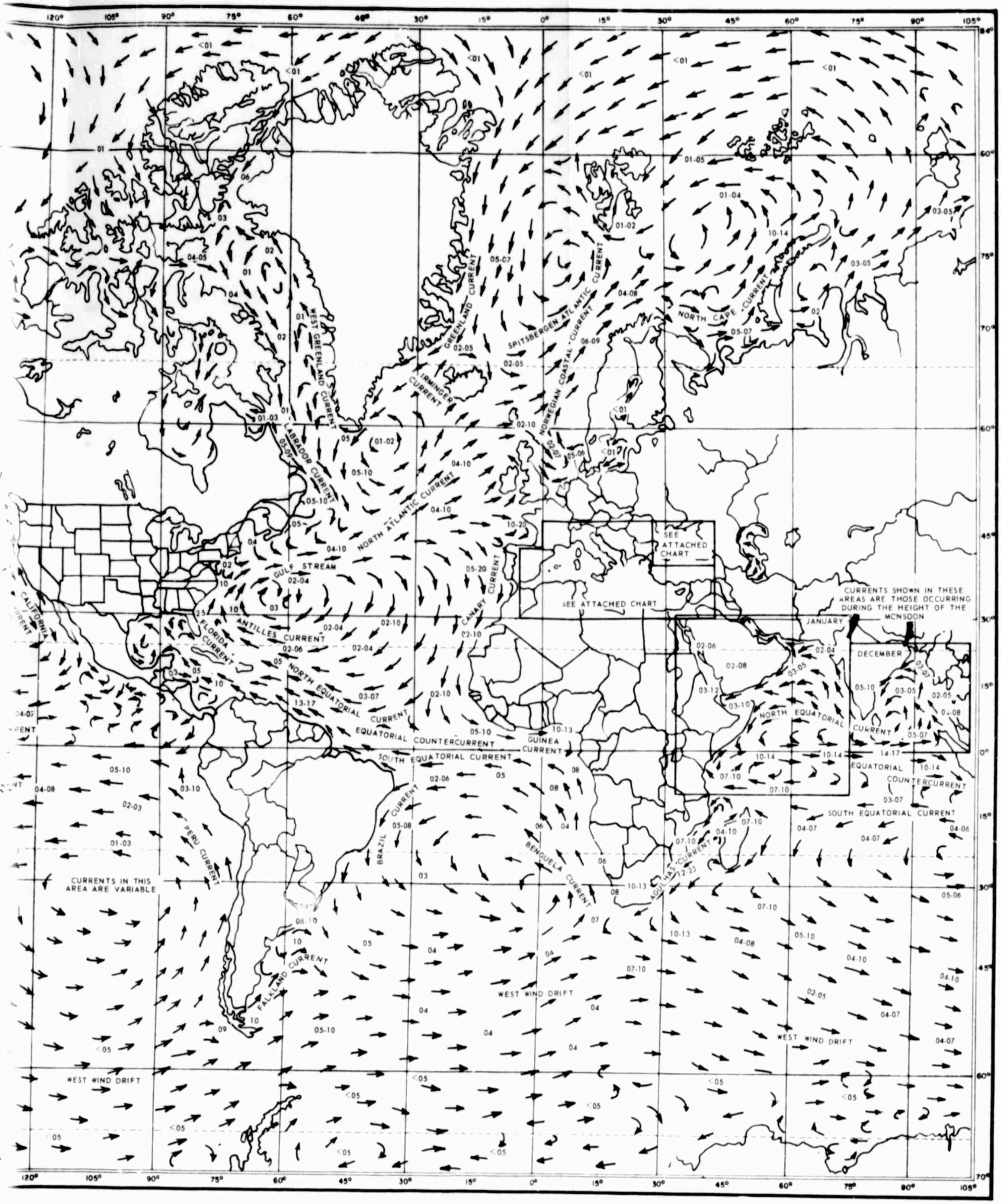
Ocean currents usually are strongest near the surface and sometimes attain considerable speed (Florida Current may attain speeds of 5 knots or more). In the middle latitudes, however, the strongest surface currents rarely reach speeds above 2 knots. At depths, currents generally are slow (less than 0.5 knot) except that recent investigations have disclosed the existence of high velocity streams, such as the Cromwell Current in the Pacific Ocean. At very great depths, in trenches for example, it was believed that currents were imperceptible, but now some evidence indicates a measurable flow.

Winds are the primary generating force responsible for the general current patterns shown in Figures 10 and 11. Figures 13 and 14 show the wind systems of the world whose characteristics are similar in many



A

FIGURE 10. AVERAGED WORLD WIDE CURRENTS, WINTER (JAN



WIDE CURRENTS, WINTER (JANUARY, FEBRUARY, MARCH)

B

LEGEND

- - —→ MOST USUAL SET OF PREVAILING CURRENT

— —→ MOST PROBABLE SET

NUMERALS INDICATE MEAN SPEED IN KNOTS.

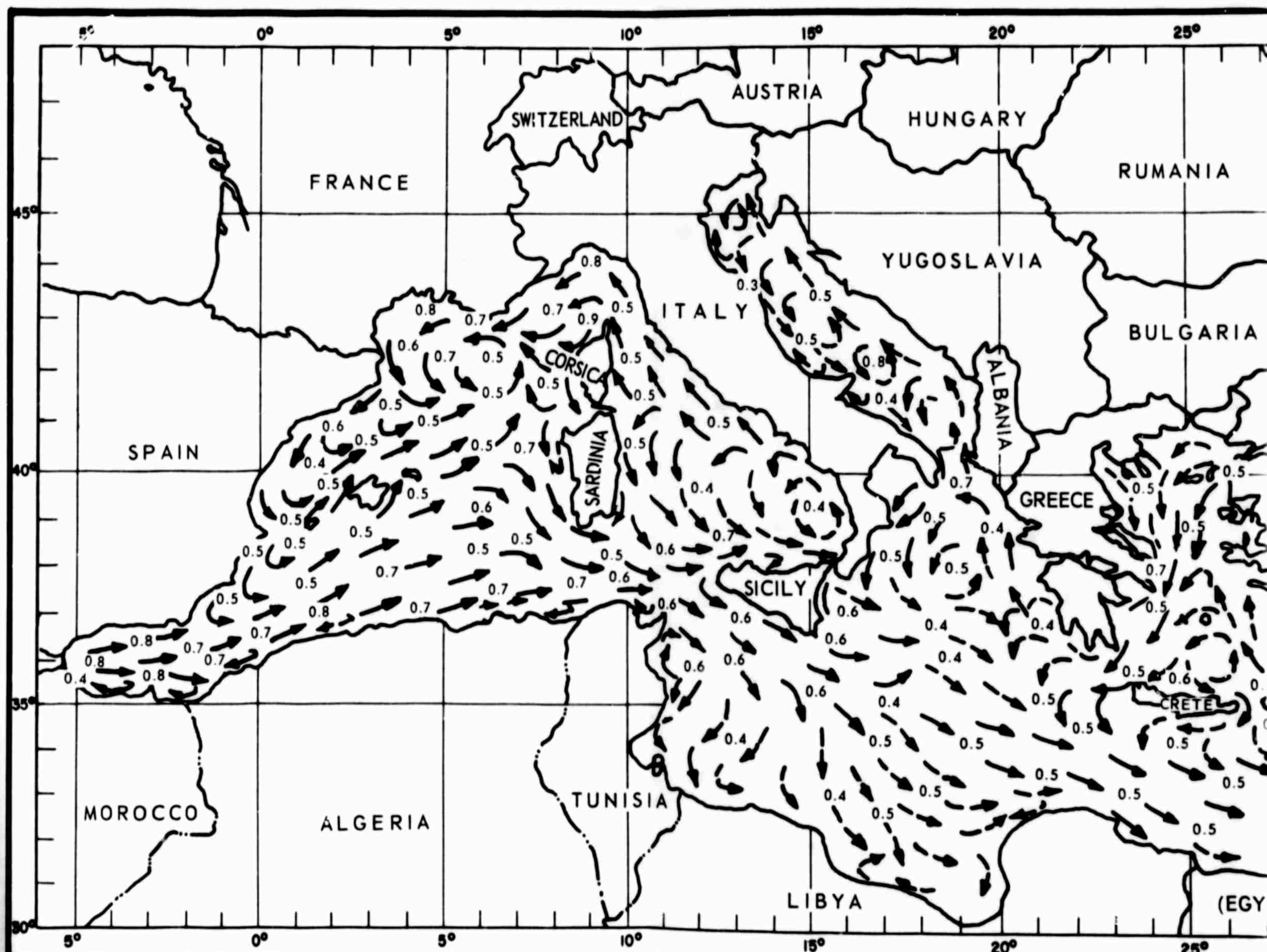
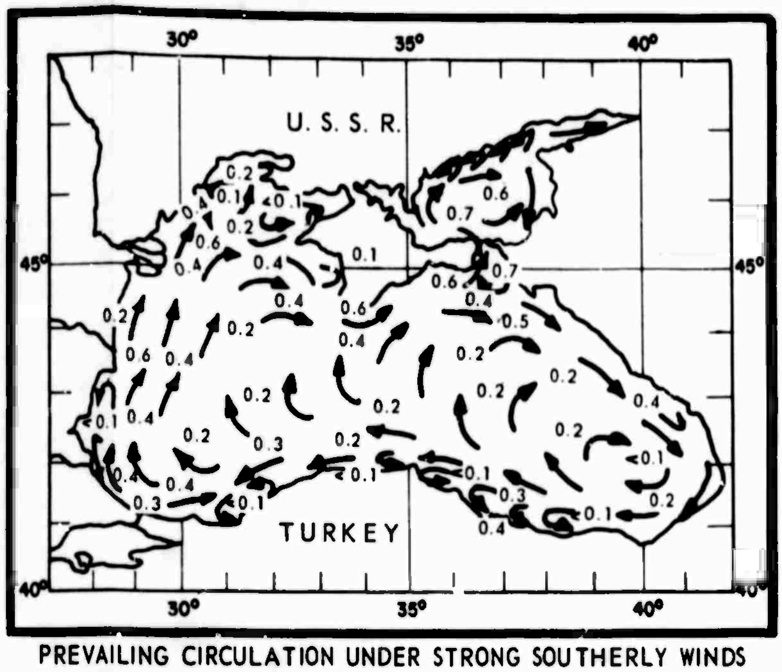


FIGURE 10A. GENERAL SURFACE CIRCULATION, JANUARY THROUGH

A



LEGEND
.ET OF PREVAILING CURRENT
..E SET
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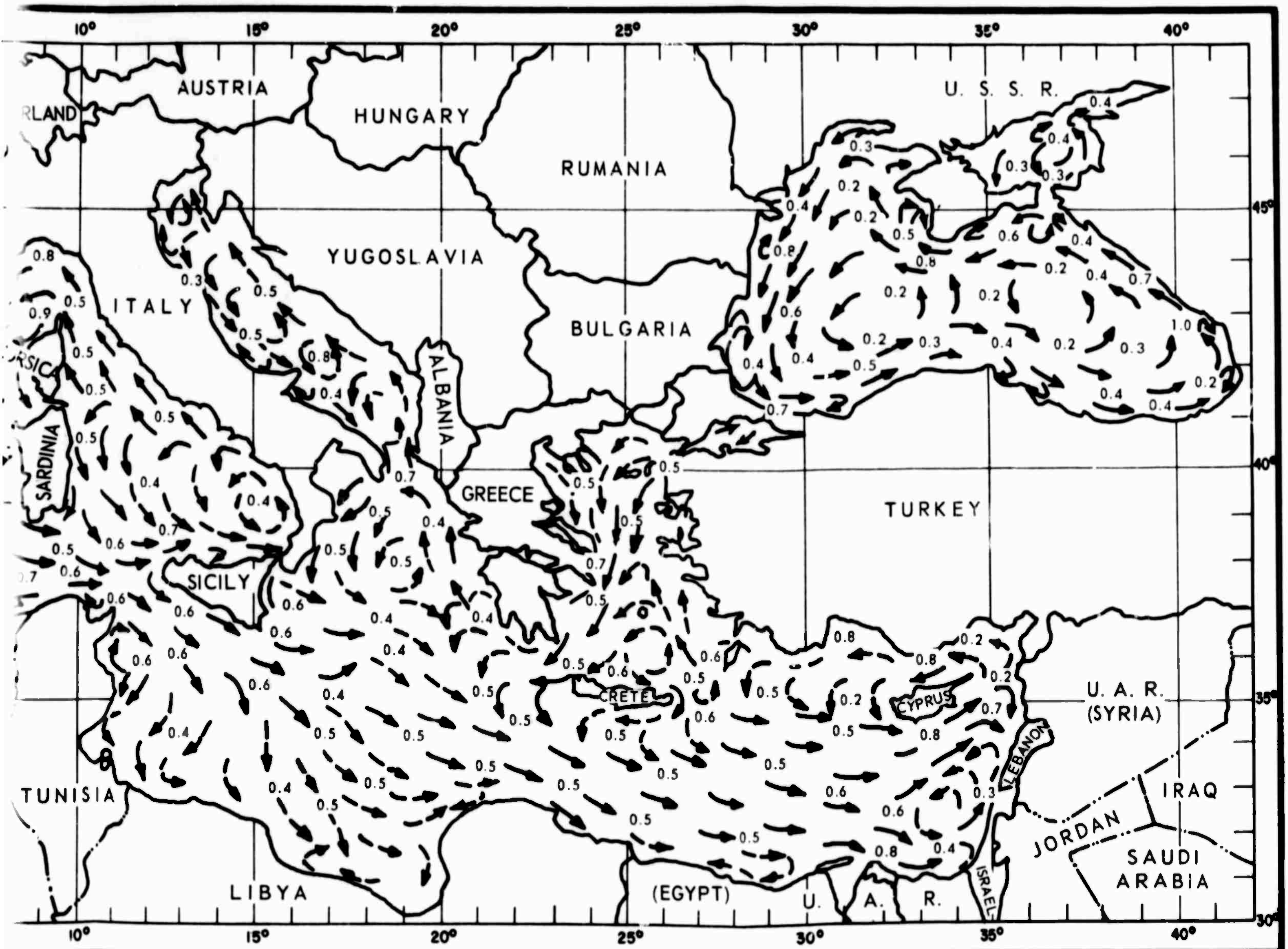
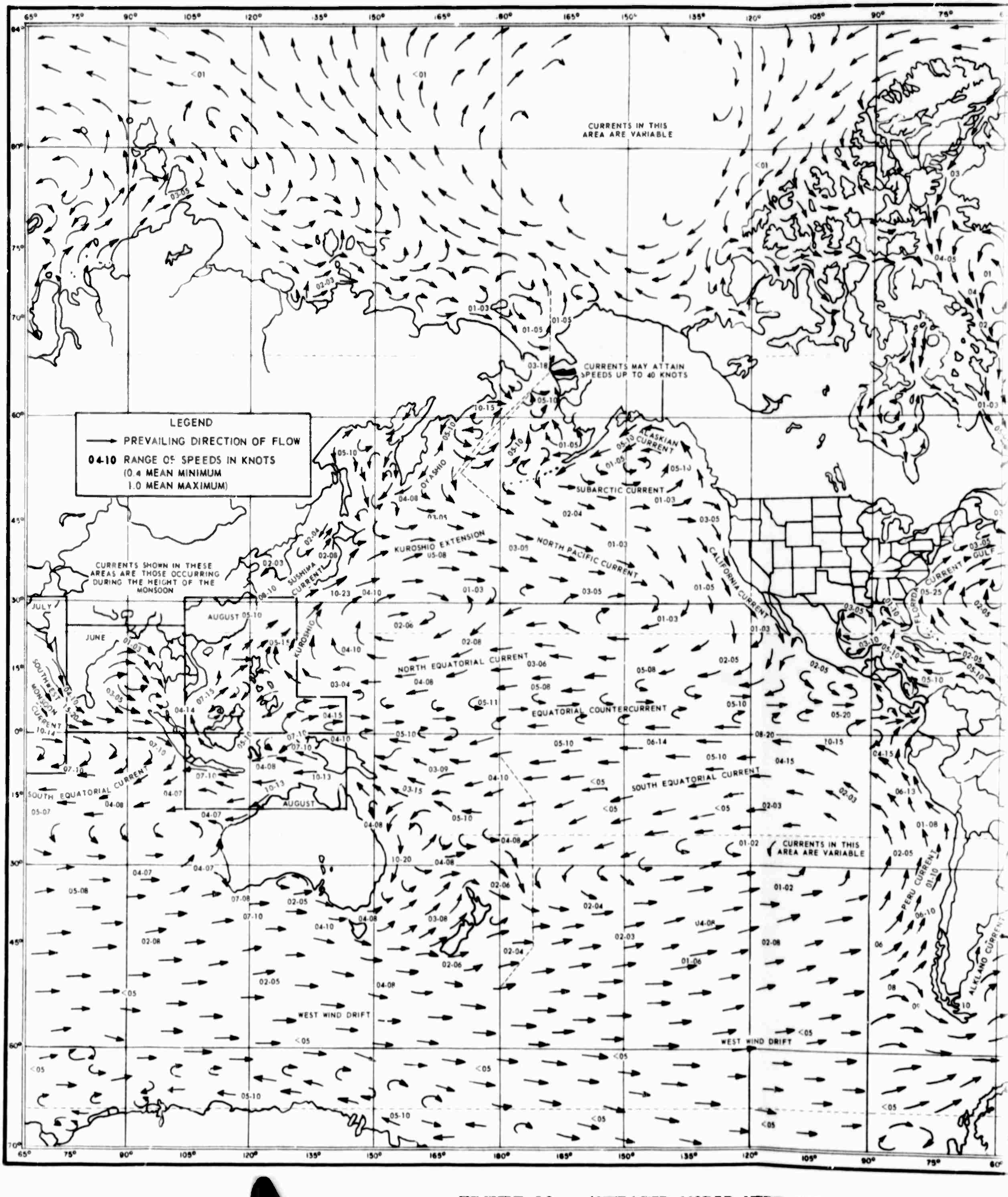


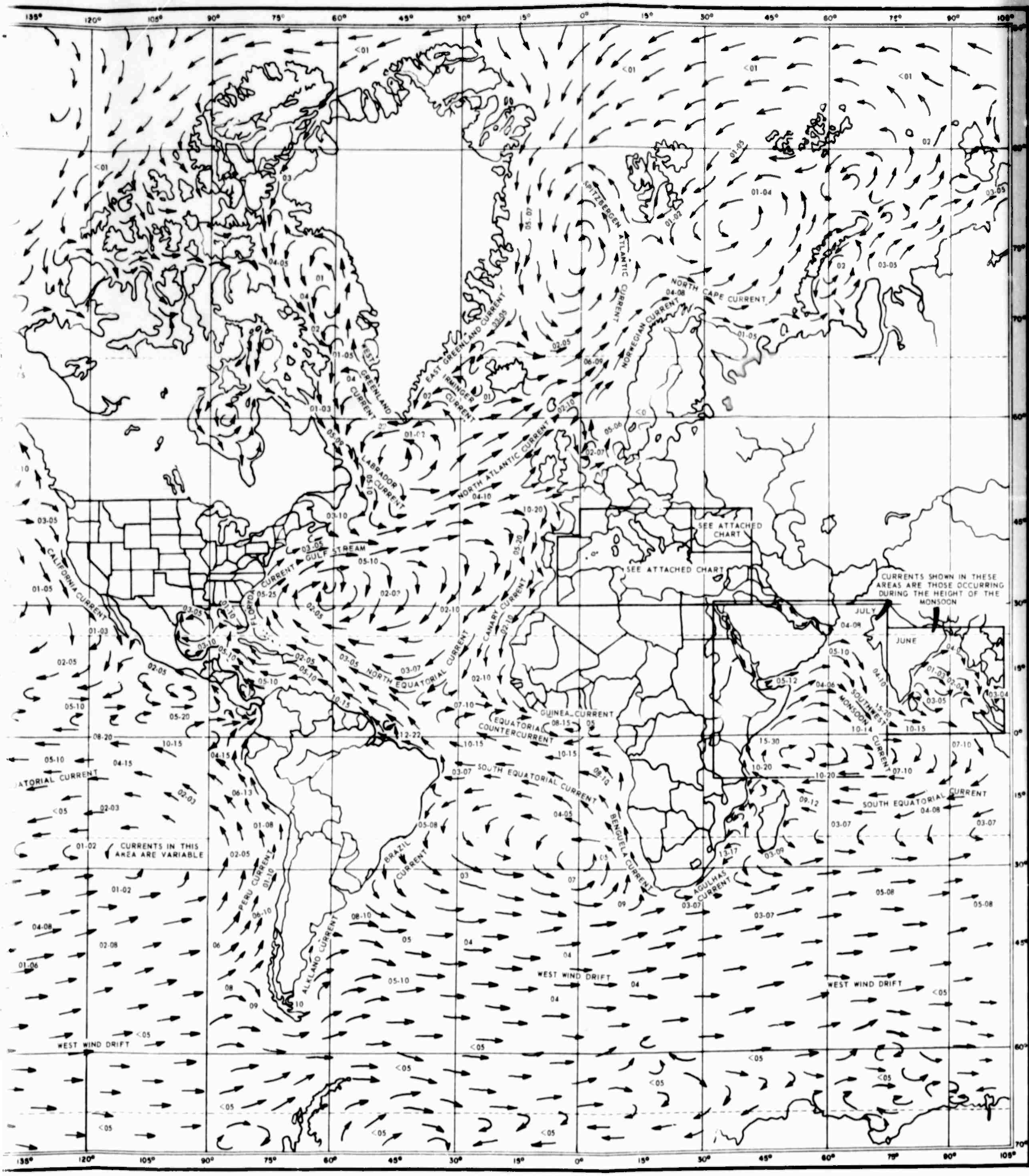
10A. GENERAL SURFACE CIRCULATION, JANUARY THROUGH DECEMBER

B



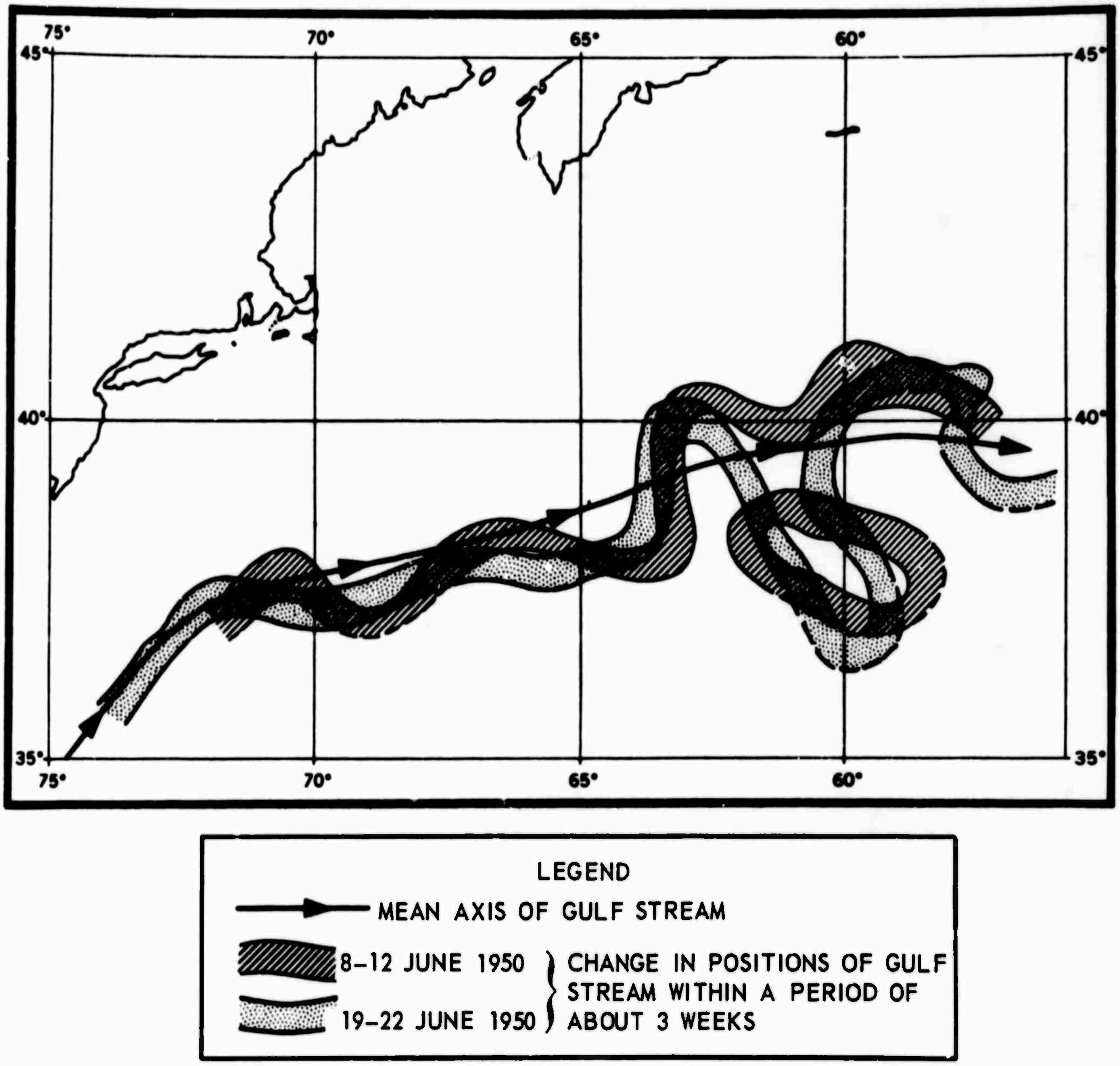
A

FIGURE 11. AVERAGED WORLD WIDE CURRENTS, SUMMER (JU



WORLD WIDE CURRENTS, SUMMER (JULY, AUGUST, SEPTEMBER)

B



(FUGLISTER & WORTHINGTON, 1951)

FIGURE 12. EXAMPLE OF GULF STREAM MEANDERINGS

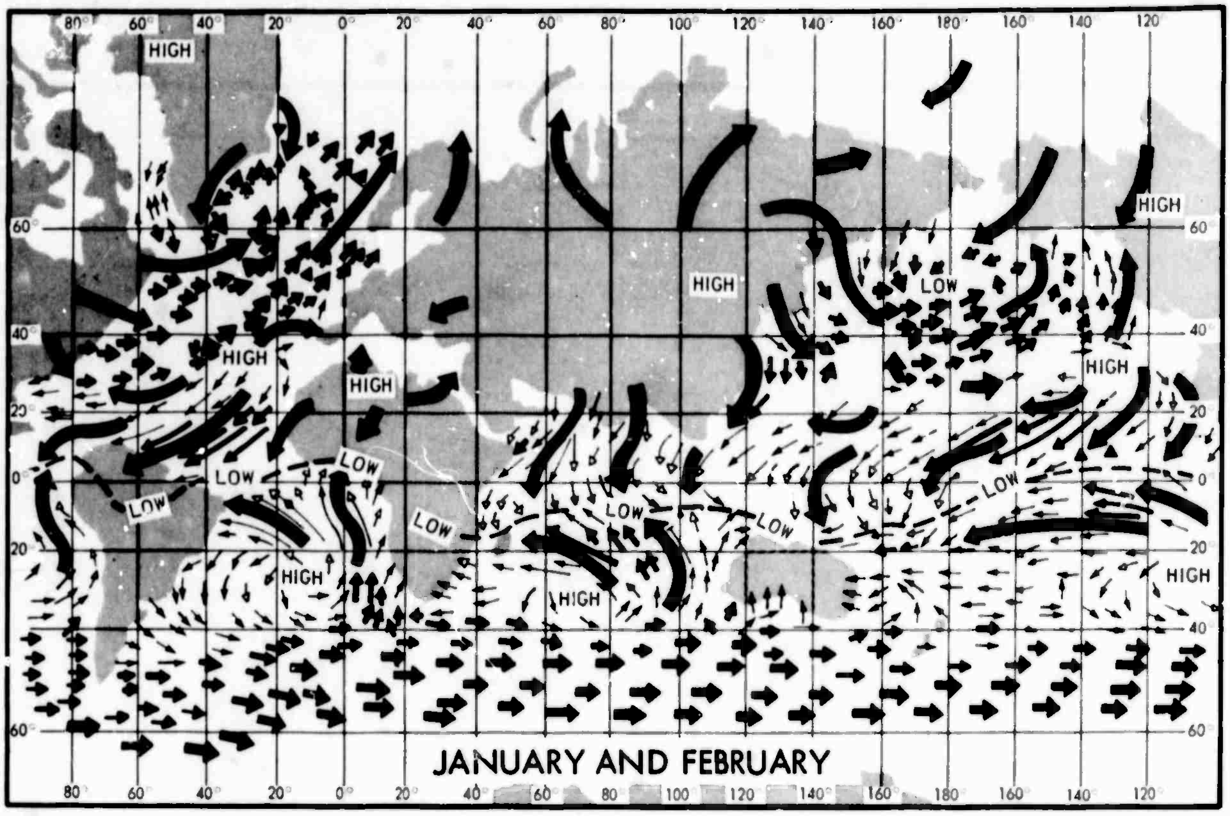


FIGURE 13 GENERALIZED PATTERN OF ACTUAL SURFACE WINDS IN JANUARY AND FEBRUARY

KEY

PREVAILING WINDS

LENGTH of arrow indicates generalized degree of CONSTANCY OF WIND DIRECTION

WIDTH of arrow indicates average FORCE OF WIND

- 20- Knots
- 15–20 Knots
- 10–15 Knots
- 10– Knots
- DIRECTION OF MOVEMENT OF AIR MASS

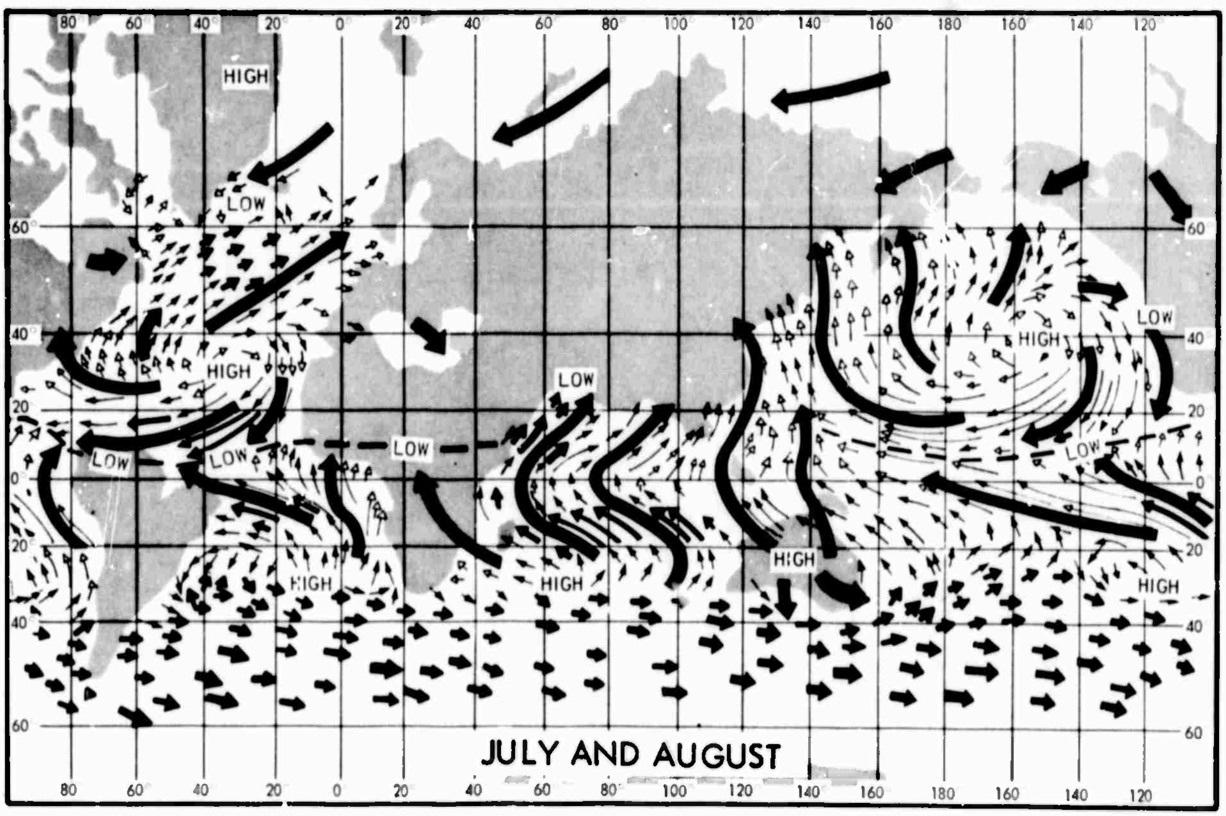


FIGURE 14 GENERALIZED PATTERN OF SURFACE WINDS

respects to the currents shown in Figures 10 and 11. A wind-driven current does not flow in exactly the same direction as the wind, but is deflected by the earth's rotation. This deflecting effect (Coriolis force), greater at higher latitudes and more effective in deep water, is to the right of the wind direction in the Northern Hemisphere and to the left in the Southern Hemisphere. At latitudes between 10°N and 10°S, the current usually sets downwind. In general, the angular difference in direction between the wind and the surface current varies from about 10° in shallow coastal areas to as much as 45° in some areas of the open ocean. The angle increases with depth of current, and at certain depths the current may flow in the opposite direction to that at the surface (Ekman Spiral). Some major wind-driven currents are the West Wind Drift in the Antarctic, the North and South Equatorial Currents that lie in the trade wind belts of the ocean, and the seasonal (monsoon) currents in the western Pacific.

Ocean currents have a definite structure which often is obscured near the surface by effects of winds and other factors. Large density gradients may develop in the vicinity of strong ocean currents. Within the prevailing current systems centered at midlatitudes, the surfaces of equal density slope downward toward the right of the observer facing downcurrent in the Northern Hemisphere and toward his left in the Southern Hemisphere. In areas where currents carrying water of different densities converge, confused currents usually exist. Examples of such areas are the convergence of the warm Gulf Stream and the cold Labrador Current in the western Atlantic, and the warm Kuroshio and the cold Oyashio in the western Pacific.

In deep water, currents are related to the thermocline (layer of rapid temperature decrease with depth), and 3 general situations may be recognized: 1) Above the thermocline the subsurface currents have about the same speed and direction as surface currents, 2) current speed frequently decreases sharply at the thermocline, and 3) below the thermocline, current speeds decrease gradually with depth, and their direction may differ appreciably from that above the thermocline. The depths to which significant subsurface flow can be detected vary with each current or in different parts of the same current. Usually, water having characteristics of that on the surface extends deeper near the axis of major current systems. Though seasonal warming may develop a sharp shallow thermocline, the stronger current and resulting turbulence near the axis of a system produces a deeper homogeneous layer.

When adjacent currents of different physical characteristics set in opposite directions, such as the cold south-setting coastal current and the warm north-setting Gulf Stream along our east coast, warm water often forms a shallow wedge over the denser colder water. For example, Figure 15 shows a latitudinal cross section of two currents flowing side by side. In this figure the vertical scale is greatly exaggerated, as the distance from A to C represents about 100 miles. The submarine in Figure 15 in proceeding from A to C travels through two areas of fairly deep surface layers near A and C. In each of these layers the thermoclines are deep. Away from the centers of these current systems the surface layer shoals, and near B (the boundary between the two currents) the warm water overlies

the cold water as a shallow wedge. Tongues of warm and cold water intertwine along this interface.

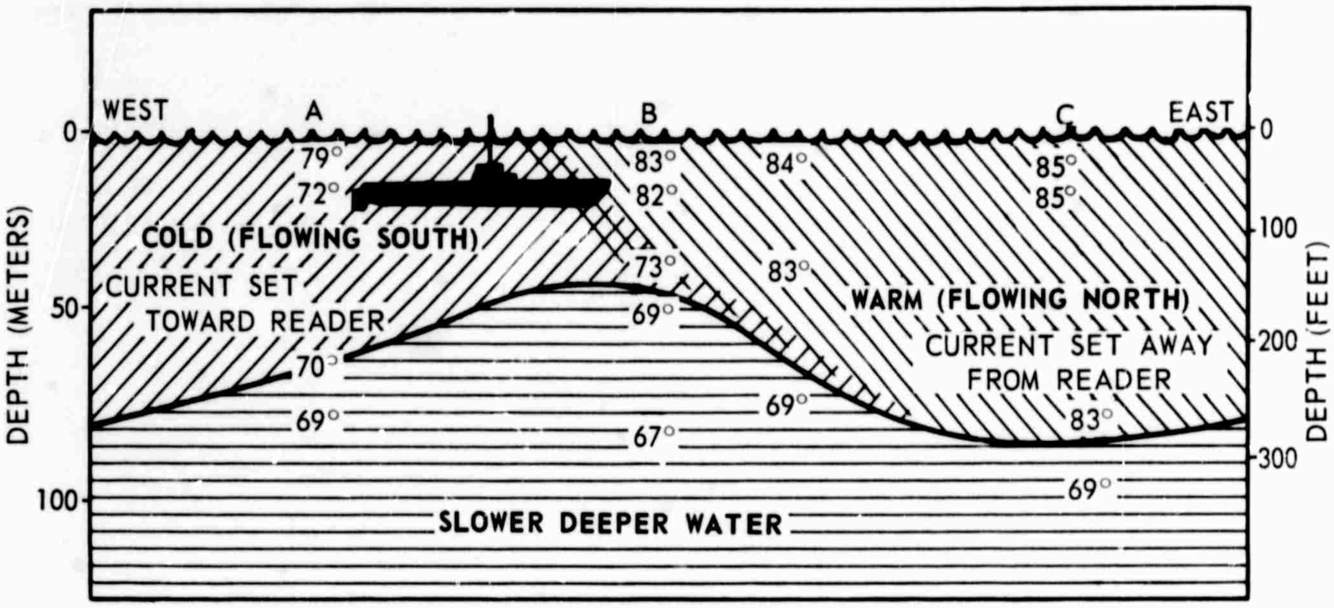


FIGURE 15. IDEALIZED CROSS SECTION OF AN OCEAN CURRENT AND COUNTERCURRENT

In addition to currents that travel horizontally in the sea, there are currents that move vertically. In some areas this vertical movement is primarily responsible for the occurrence of strong thermal gradients. Most vertical circulation occurs in the upper 300 meters. Vertical circulation at great depths is represented by a general overturning of the water masses, rather than vertical transport by definite currents.

A highly variable phenomenon much influenced by local water movements and weather is the vertical downward movement (sinking) of heavier surface water. Sinking may occur in almost any part of the ocean, but ordinarily occurs at the convergence between two water masses of different characteristics, especially where there is considerable cooling or evaporation of surface water. Such a convergence area may have the choppy appearance of a tide rip and contain debris brought into it by the converging currents. Also, the downdraft sometimes is strong enough to sink debris of slight buoyancy, and the surface currents may be strong enough to hold a drifting vessel in the convergence despite crosswind. Unfortunately very little information is available about the size and form of the downward currents.

Converse to the descending movement of surface waters is the ascending movement (upwelling). Upwelling may occur anywhere in the ocean, but most often occurs in coastal regions where a strong wind blows the surface water seaward and allows the deeper colder water to rise in its place. Upwelling normally occurs along the west coasts of continents, being particularly strong off Peru, California, and the west coast of Africa. Offshore winds tend to produce upwelling, but wind paralleling the coast also may cause

upwelling along the shore as Coriolis effect will deflect the flow of water considerably (Fig. 16). For example, a north or northwest wind along the coast of California causes upwelling, whereas a southerly wind does not. In the Southern Hemisphere southerly winds may cause upwelling, as off the coast of Peru.

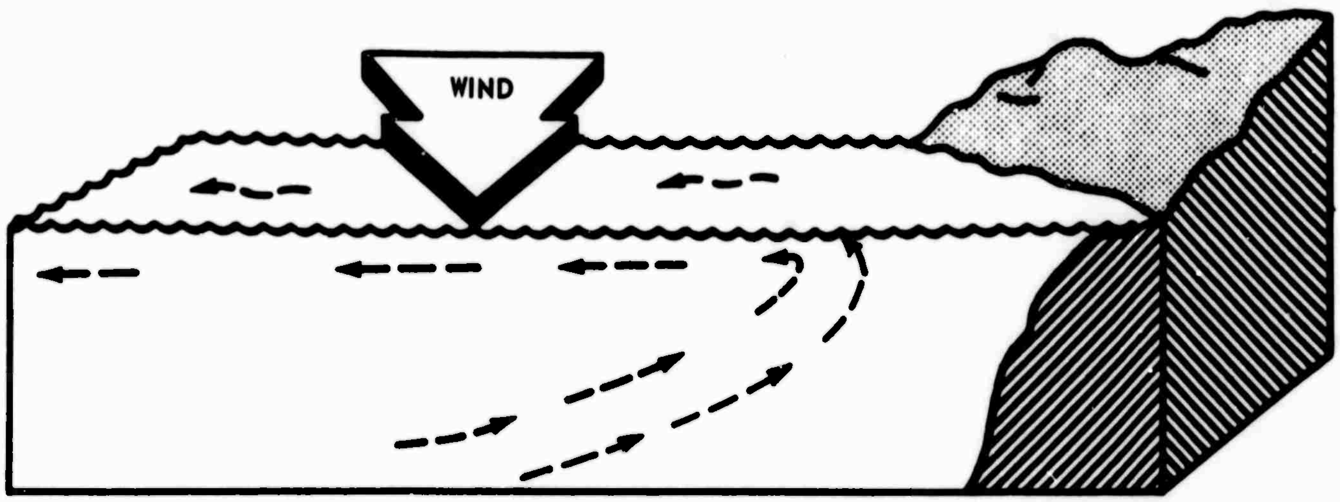


FIGURE 16. EXAMPLE OF UPWELLING IN NORTHERN HEMISPHERE

Eddies, which vary in size from a few miles to 75 miles or more in diameter, branch from major currents. Large eddies are common on both sides of the Gulf Stream from Cape Hatteras to the Grand Banks. How long such eddies persist and retain their characteristics near the surface is not well known, but large eddies near the Gulf Stream are known to persist longer than a month. The surface speeds of currents within these eddies when first formed may reach 2 knots. Smaller eddies have much less momentum and soon lose their surface characteristics through wind mixing.

Much of the theory regarding open ocean currents does not apply to water movement shoreward of the 100-fathom curve. For example, in open coastal areas beyond the influence of the bottom, tidal currents are rotary in nature, that is, the direction changes progressively clockwise (in the Northern Hemisphere) around the compass in one tidal cycle with the tidal flow extending to the bottom at decreasing speeds; whereas, nearshore tidal flow becomes oscillatory.

MARINE GEOLOGY

Knowledge of submarine geology and geophysics has become increasingly important to all concerned with undersea warfare. Specifically, this knowledge is important when considering problems in: 1) Transmission of underwater sound, 2) concealment of submarines, 3) false sonar targets, and 4) submarine navigation. Although more field surveying remains to be done in order to fill in the many gaps in our knowledge of submarine geology and geophysics, an understanding of what is known today will go

a long way toward helping to solve the many problems encountered in, for example, sonar operation.

The unconsolidated sediments found on the bottom can be classified in a number of ways. There are two methods in general use. The first is a modification of that proposed by C. K. Wentworth. It associates ranges in partical size with a descriptive term as shown in the following table:

Sediments	Diameter of Grain Size (millimeters)
Boulder (rock)	Greater than 256.0
Gravel:	
Cobble	64.0 to 256.0
Pebble	4.0 to 64.0
Granule	2.0 to 4.0
Sand:	
Very coarse	1.0 to 2.0
Coarse	0.5 to 1.0
Medium	0.25 to 0.50
Fine	0.125 to 0.250
Very fine	0.0625 to 0.1250
Silt:	
Coarse	0.0313 to 0.0625
Medium	0.0156 to 0.0313
Fine	0.0078 to 0.0156
Very fine	0.0039 to 0.0078
Clay:	
Coarse	0.00195 to 0.00390
Medium	0.00098 to 0.00195
Fine	0.00049 to 0.00098
Very fine	0.00024 to 0.00049
Colloids	less than 0.00024

A second system is not one of quantitative classification, but rather it defines areas of bottom sediment descriptively, as explained below. A typical example of bottom sediment depiction is shown in Figure 17. Since a sediment rarely occurs which is entirely within one Wentworth size grouping, generalized descriptions are used. Thus, sand, for example, may consist of material predominantly of sand-size material (according to Wentworth) but in addition may contain minor amounts of material of other

sizes. When the minor amounts of other size material increase to an appreciable amount (>20%), the mixture can be classified and shown separately on the chart as such. For example, some areas of sand may contain small amounts of mud but are still shown as sand. Other areas having a mixture of mud and sand, each in appreciable amounts, are designated as mud-sand.

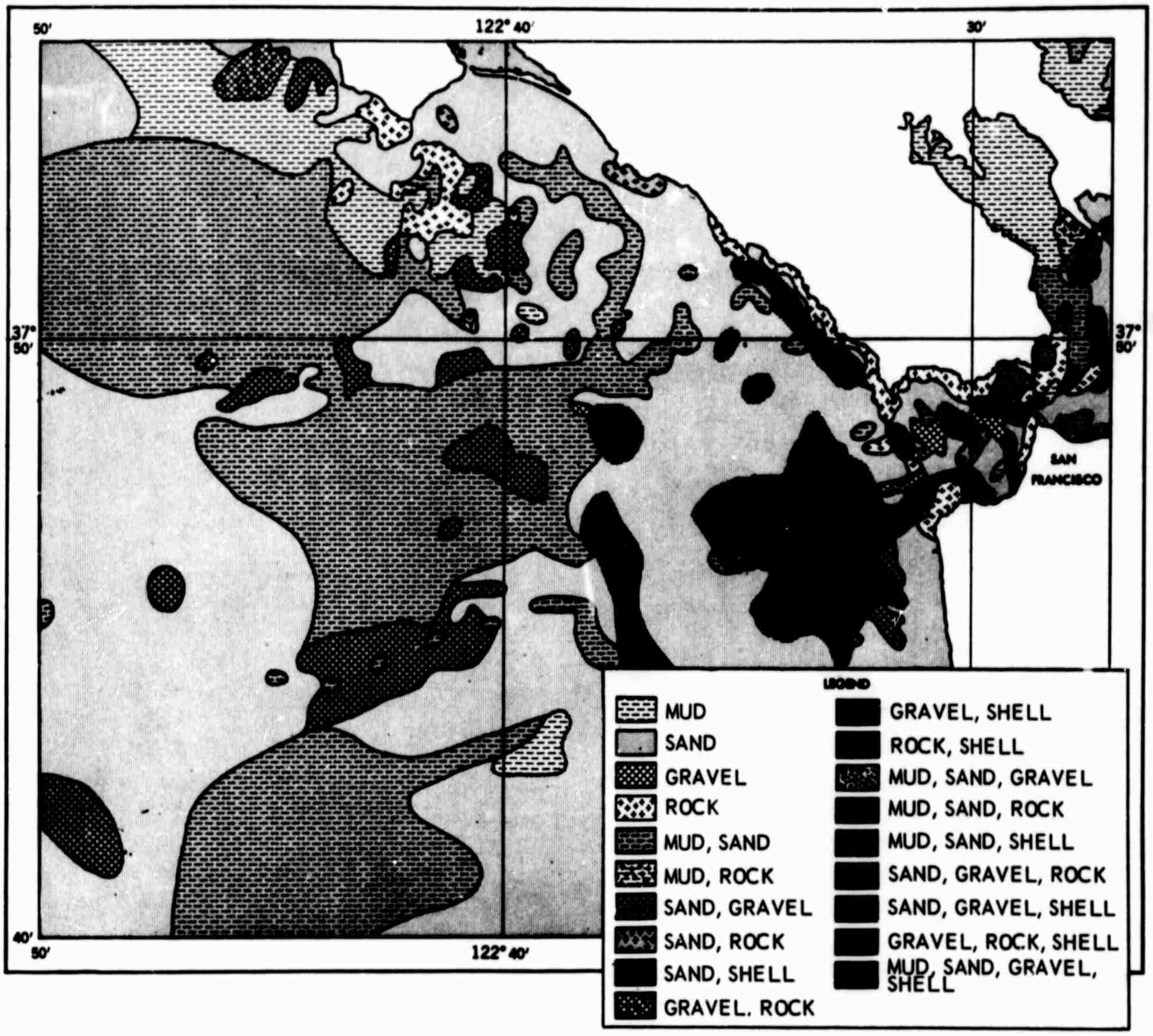


FIGURE 17. CHART OF BOTTOM MATERIALS OFF A NONGLACIATED COAST

The materials of the sea bottom range in size from the finest clays and colloids to bedrock; most of the particles are from clay to sand in size. The photograph of the bottom, Figure 18, gives an indication of the wide variation that occurs. The type of bottom in an area is dependent on factors such as the geologic history of the area and the present

environmental conditions. The composition of the bottom may include bed-rock, living or dead coral, calcareous algae, or detrital material.

FIGURE 18. BOTTOM PHOTOGRAPH

Detrital material in deep water refers to particles that have comparatively short geologic history since separation from their parent rock, coral, or algal masses. This material has been transported from the parent bodies distances ranging from few to many miles. The transporting medium is usually water currents, although minor amounts of material have been transported by the atmosphere before being dropped into the sea. As long as the water masses move at sufficient velocity, certain sizes of particles remain in suspension and can be transported from one place to another. However, when the velocity drops below a critical value, some of the larger particles in suspension are deposited while the finer material remains in suspension. If the velocity continues to decrease, only the smallest particles remain in suspension. Thus, the size of the material being transported and deposited depends on velocity and associated turbulence of the water.

Marine sediments are subdivided into two major groups, pelagic and terrigenous. Pelagic deposits are those found in deep water far from shore and may be either organic or inorganic in origin. These deposits are characterized by clay-sized particles and a paucity of terrigenuous materials. The most common constituents are clay minerals and the remains of planktonic unicellular organisms. Pelagic deposits containing more than 30% tests are known as oozes. Usually these sediments are white or very light colored. A popular misconception is that the oozes are very fine grained, but often they are in the coarse to fine sand range. Oozes can

be subdivided into two general types, calcareous or siliceous, depending on the composition of their organic constituents. These two types can be further subdivided according to the characteristic organism present, so that within the calcareous category there are globigerina, pteropod, and coccolith oozes, and within the siliceous category there are diatom and radiolarian oozes.

Pelagic inorganic deposits, known as red clay, contain less than 30% organic material. Although called red clay, since the first samples collected were brick red, these deposits also occur in other colors; in fact most of them are brown or buff colored. Regardless of color, these deposits are composed predominantly of material of clay size or smaller.

Terrigenous deposits are found near shore and in most semi-enclosed seas, gulfs, and bays; they generally contain much material of land origin. Terrigenous deposits cover a wide range of depths and a great variation in color, texture, and composition. The classification of these sediments is not as straightforward as that of the pelagic sediments. A number of systems have been suggested, but the character of terrigenous deposits depends so much upon local conditions that no one system has a very wide application. Compared to pelagic deposits, terrigenous deposits cover a small portion of the sea floor. Since they show great changes in characteristics within relatively short distances seaward, the occurrence of transitional types makes definition difficult. Descriptive terms are used which depict the color, texture, and composition of the material; such terms include: "green diatomaceous silty clay" or "gray calcareous sand."

BOTTOM TOPOGRAPHY

Our present knowledge of bathymetry (submarine topography) is based on soundings taken over many centuries. Early observations usually consisted of sounding only in relatively shallow water where dangers to navigation were the prime concern. Deepwater soundings were of little concern until the laying of the first transoceanic cables in the late 19th century.

The early soundings were made with hemp rope, which made the task long and tedious; a single sounding could take several hours. Improvements over the years included first the use of wire rope, then single-strand wire for deep soundings. With the development of echo-sounding equipment in 1922 the picture changed completely. This innovation made a deep sounding possible in only a few seconds by measuring the travel time of a sound impulse to the sea bottom and return (assuming a mean vertical sound speed of 1463 m/sec). An improvement of this technique was the development of automatic echo-sounding equipment which produced a continuous trace of the profile of the sea floor over which a ship was passing. With this type of information accurate bathymetry charts can be made.

The ocean bottom is considered to consist of 3 major provinces: 1) the continental shelf, 2) the continental slope and rise, and 3) the ocean basin. A typical bottom relief profile is shown in Figure 19.

In general the continental shelf slopes gently seaward to depths usually from 60 to 100 fathoms and terminates at its seaward extremity where the gentle slope (1° or less) becomes much steeper (1° to 4°). This

break ranges from a sharp edge to a rounded shoulder. Continental shelves vary greatly in width and slope. For example, off glaciated coasts and areas which are predominantly low and flat the shelf may be very wide, whereas off mountainous coasts the shelf may be virtually absent. World-wide, the shelf width averages about 30 miles, with a range from 0 to over 800 miles. Although the average slope of the shelf is gentle within the shelf boundaries, terraces, ridges, hills, depressions, and steepwalled canyons may occur. Where these features provide extremes in relief, normally not recognized as being within the shelf province, the area is called a continental borderland. One of the well-known examples of a borderland is the area off the southern California coast. Instead of the usual gently sloping area of minor relief typical of shelf areas, this area is an extensive region of peaks and valleys more or less parallel to the topographic trends of the adjacent land. Some of these peaks extend above sea level to form islands, and some of the valleys reach depths exceeding 800 fathoms (one as much as 1,400 fathoms). All of these features occur within an area that normally would be designated a continental shelf.

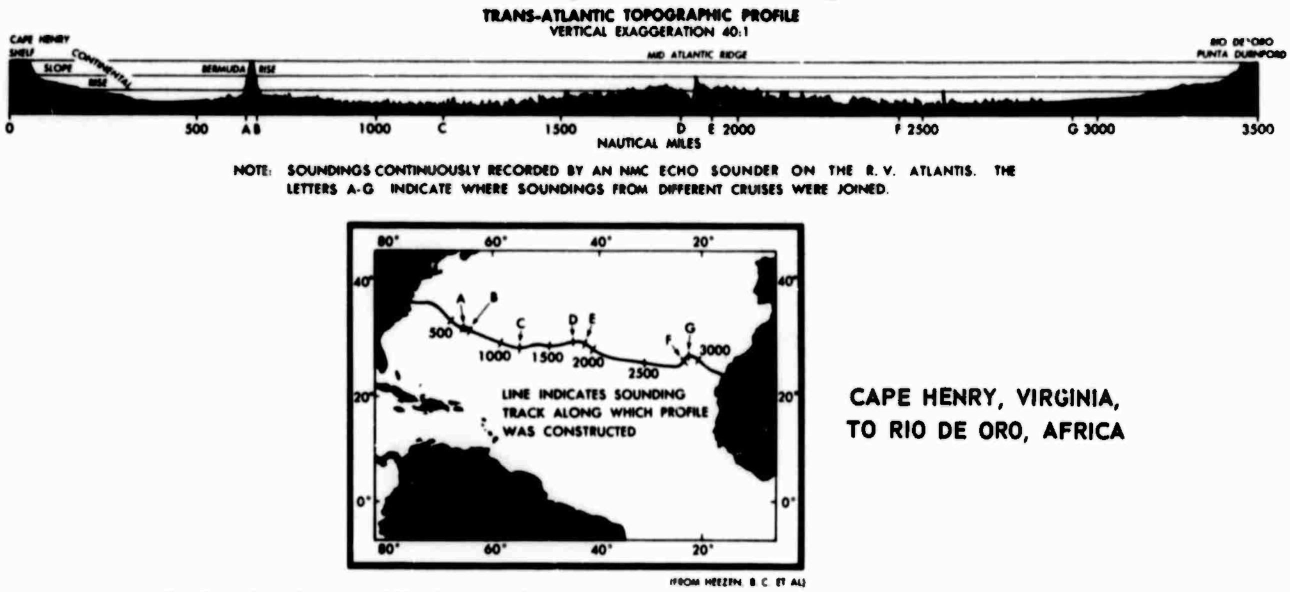


FIGURE 19. BOTTOM PROFILE OF THE NORTH ATLANTIC OCEAN

Beyond the break at the seaward edge of the shelf is the continental slope which inclines downward toward the ocean depths, giving way to a less steeply sloping zone known as the continental rise before the deep sea floor is reached. Generally the continental slope off mountainous coasts has an incline of about 1:200 but off coasts with wide, well-drained plains these slopes incline about 1:30. Extreme slopes, such as off volcanic islands where the shelf usually is absent, may incline as much as 1:11. As on the continental shelf, the slope can and usually does have minor relief superimposed. What is called gullying on land and submarine canyons in the ocean is a feature that probably is typical of most of the slope areas of the world, particularly on the upper parts of the

slopes. Most canyon heads are at or near the break in the shelf, but a few extend across part of the shelf. The Hudson River Canyon off New York is a classic example of a submarine canyon.

Continental slopes grade smoothly to the incline of the continental rise, and this in turn grades almost imperceptibly (at depths of 500 fathoms or greater) to the floor of the ocean basins.

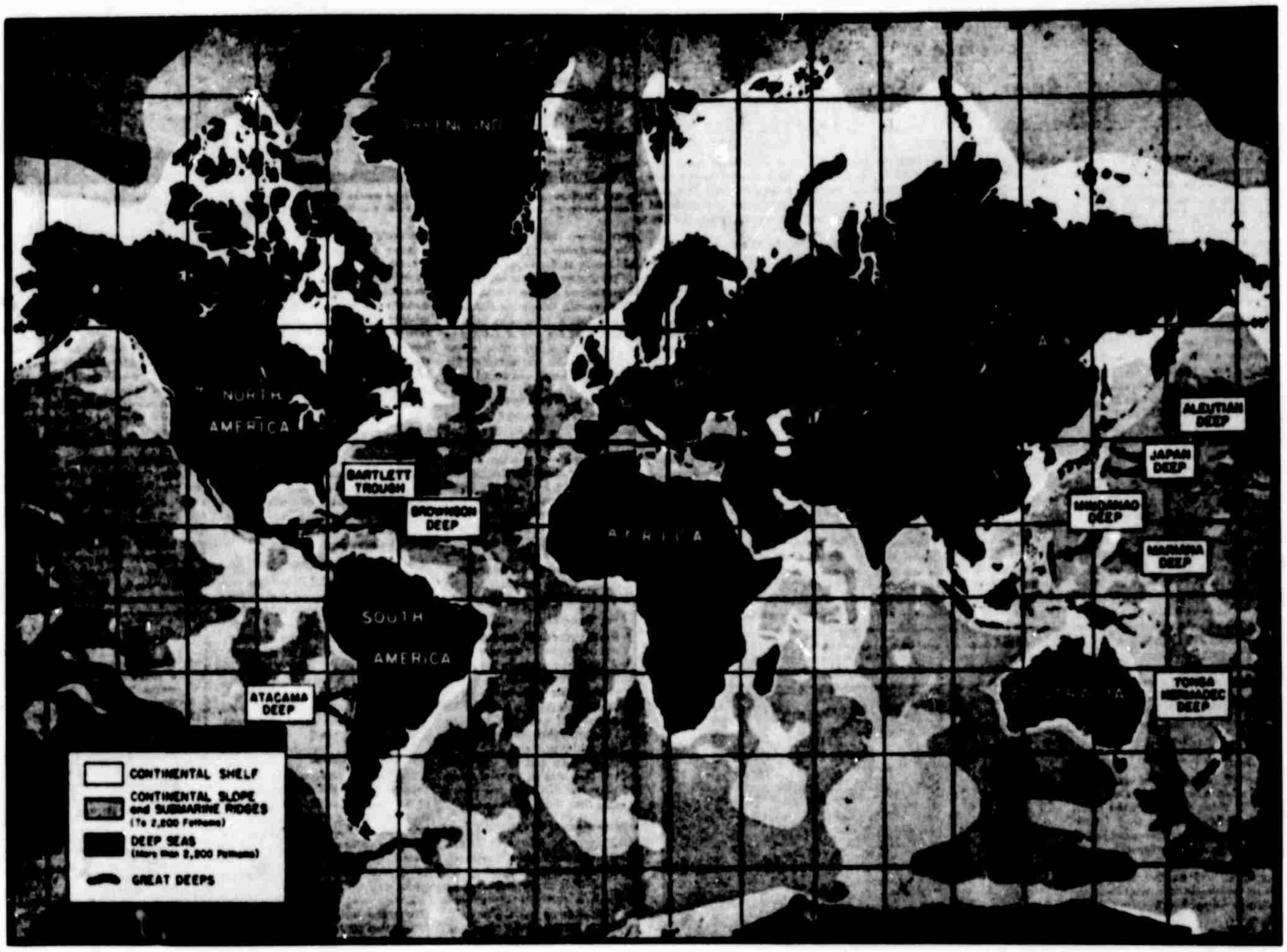


FIGURE 20. BOTTOM FEATURES

The deep sea floor is believed by many to be a flat featureless abyssal plain. Actually it does have, at most, a very gentle incline of about 1:90°. The relief superimposed on this average incline may be at least as rugged as the larger topographic features found on land. For example, seamounts rising from the ocean floor are widespread in the Pacific, and great mountain chains or ridges and deep trenches or furrows are common in all oceans. Figure 20 shows some of the major bottom features in the oceans. Figure 21 shows diagrammatically the terminology used in describing sea floor topography. The larger ridges of the bottom often are oriented parallel to the coasts of the continents so that the oceans are divided into elongate basins. Transverse ridges in turn tend to subdivide

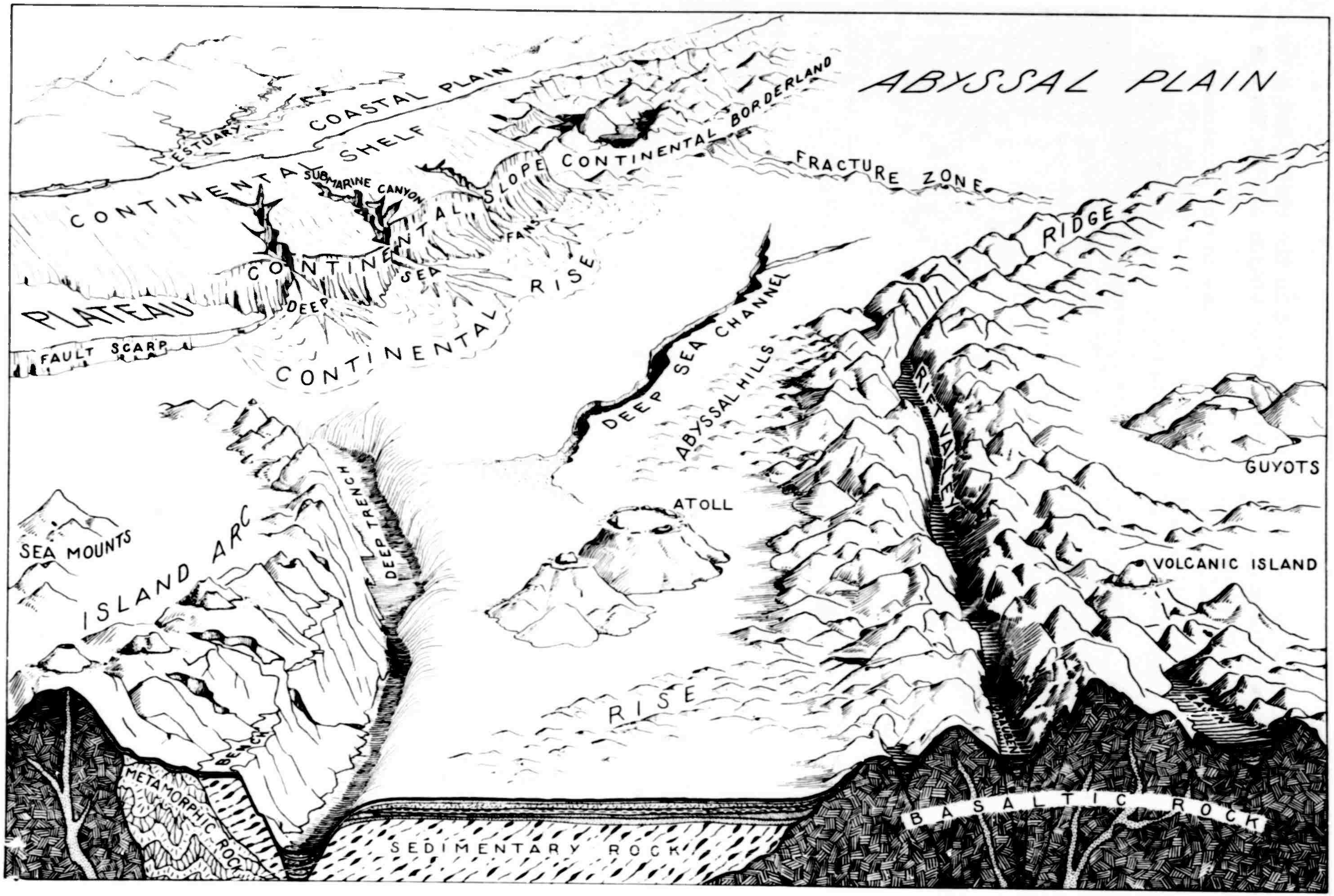


FIGURE 21. NOMENCLATURE OF UNDERSEA GEOPHYSICAL FEATURES

these basins into smaller basins. These bottom features are found in the Atlantic, Arctic, and Indian Oceans and in the western Pacific Ocean. The following are some of the major longitudinal ridges in the world:

(1) The Mid-Atlantic Ridge divides the Atlantic Ocean into the western and eastern basins. The ridge rises from depths of about 2,500 fathoms, is continuous at depths of less than 1,500 fathoms over the greater part of its length, and in several places rises above sea level to form islands such as the Azores, St. Peter and St. Paul Rocks, Ascension, and Tristan da Cunha.

(2) The Mid-Indian Ridge extends from India to Antarctica and is wider but does not rise as close to the surface as the Mid-Atlantic Ridge.

(3) In the Pacific Ocean the ridges are less conspicuous although the West-Pacific Ridge, which is composed of several shorter ridges, can be traced from Japan to Antarctica. Another ridge extends south and west of Central America.

(4) In the Arctic Ocean the Lamonosbv Ridge extends from northern Greenland to the shelf off the New Siberian Islands and separates this ocean into two basins.

Trenches are of much more limited extend than the great ridges, but within these relatively small areas the deepest points in the ocean are found. Depths in some of the deepest trenches are in excess of 10,000 meters, with the deepest sounding (11,034 meters reported by the VITYAZ) recorded in the Marianas Trench. Conversely the highest mountain on land, Mount Everest, has a height of only 8,888 meters.

Many of the sea bottom features probably have existed essentially unchanged since their formation, because little or no erosion takes place at the depth in which they are found. Sediments subject to active erosion are those near enough to the surface to be affected by the leveling action of waves or strong currents. The topography of the bottom is modified primarily by sedimentation. Sedimentary debris accumulating slowly on the bottom masks minor irregularities by filling the depressions.

Along the margins of continental shelves and on the upper portions of continental slopes, deposition from various sources can build up sediments rapidly. These sediments can become unstably situated and begin to slide. When this occurs, a submarine landslide ensues that may sweep down the continental slope and out over the continental rise and even the abyssal plain as a turbidity current. This occurrence is not predictable, and the speed of the turbidity current may vary from several knots to as much as 70 knots. The resultant layered deposits in deep water consist of coarse unsorted material foreign to the deep water area.

Figure 22 shows a series of bottom photographs taken while drifting across an unusual outcrop at about 1,800 meters in the Tongue of the Ocean (Bahamas). Figure 23 is a contour map of part of the area. The irregular features indicate the presence of currents sweeping the area clean of sedimentation but not strong enough to cause erosion. Figure 24 shows ripple marks in the sediment at a depth of about 1,929 meters. Calculations on these ripples indicate a current speed at the bottom of

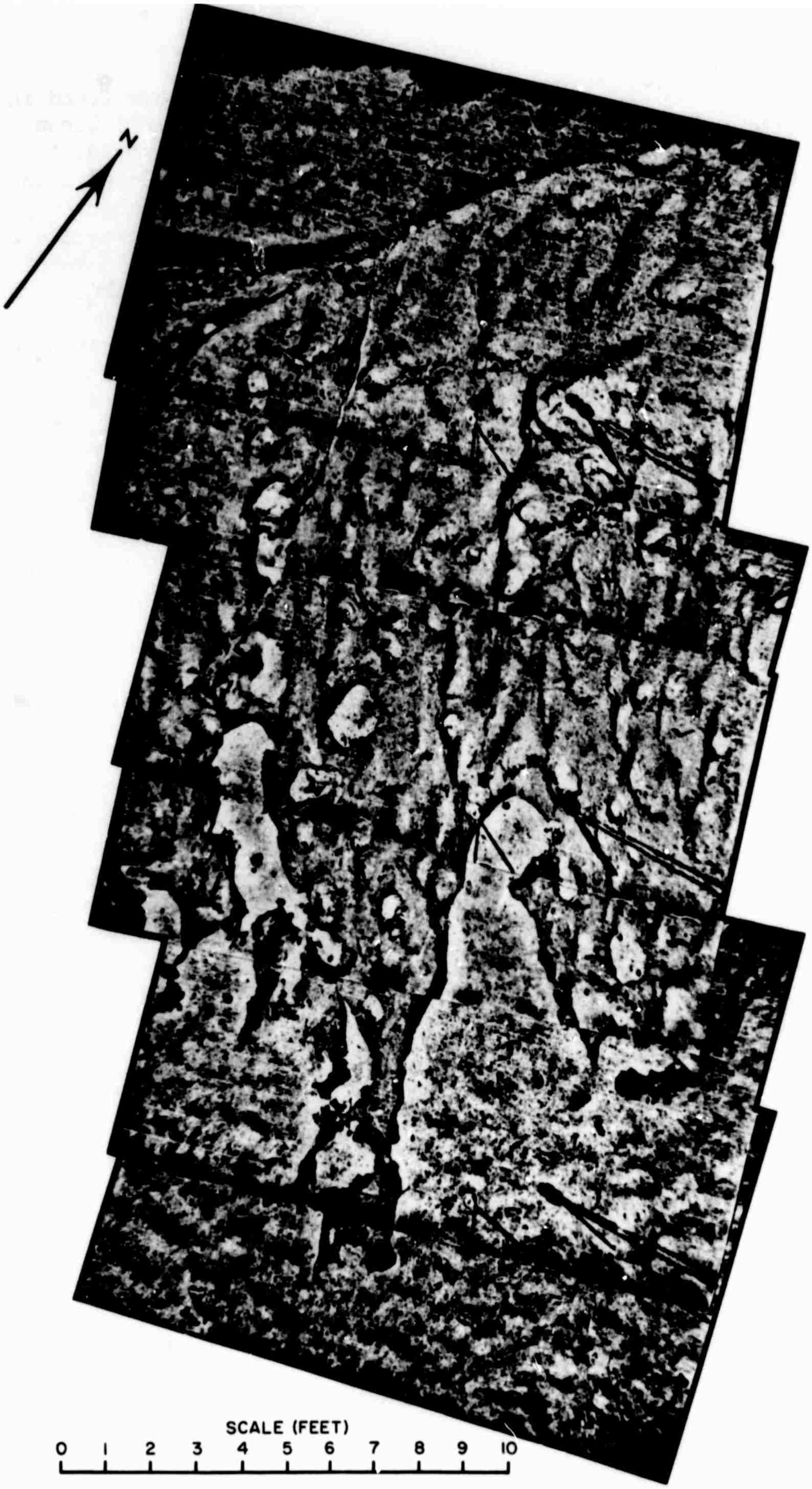


FIGURE 22 MOSAIC OF AN OUTCROP AS PHOTOGRAPHED

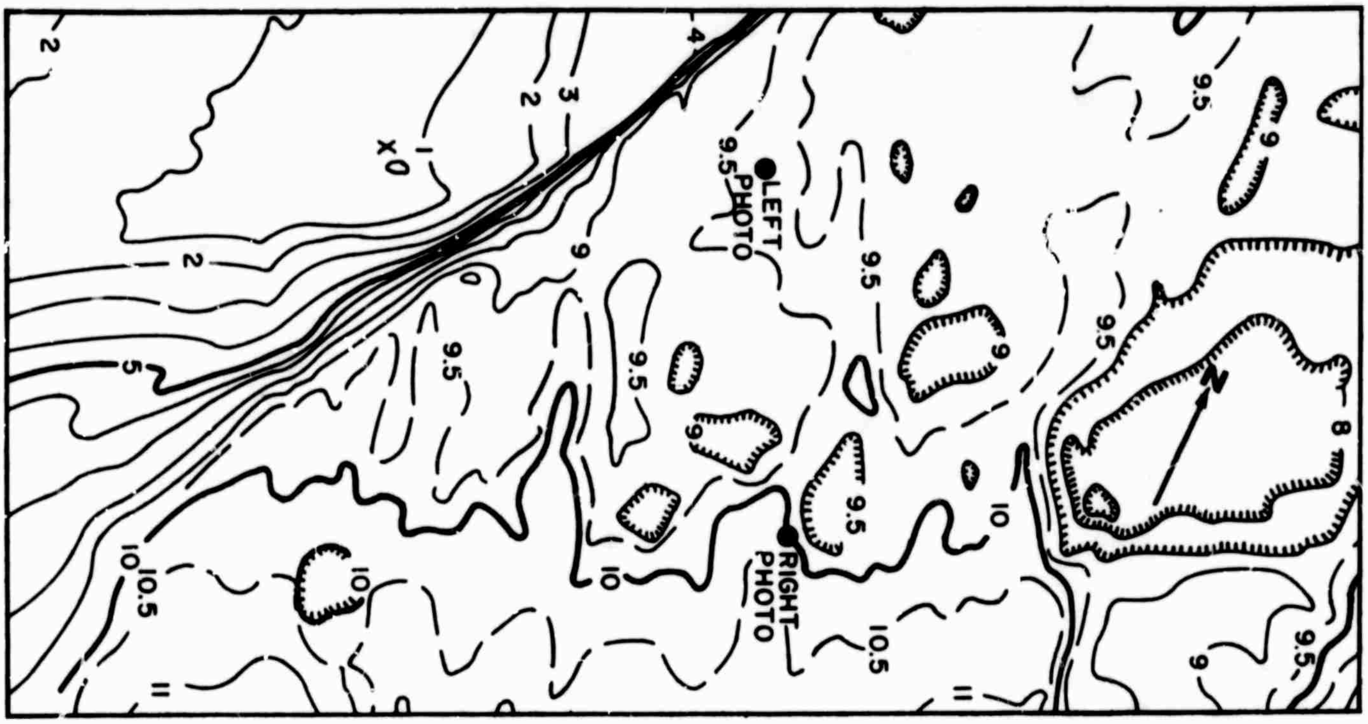


FIGURE 23. MICROTOPOGRAPHIC CONTOUR MAP OF OUTCROP. CONTOUR INTERVAL: 1 DECIMETER, SCALE: 1:17.8

between 28 and 90 cm/sec. Current speeds were sufficient to move the sediment grains and orient them into ripple formations, but not so great as to erode all features.

MARINE BIOLOGY

Fouling - Fouling is the attachment and growth of marine animals and plants upon submerged objects. Adverse effects of fouling include obstruction of salt water piping systems, dip of buoys, attenuation of signals transmitted or received by acoustic devices, and destruction of protective coatings allowing increased corrosion of metals.

The fouling community is the total accumulation of all organisms that are present on an exposed surface, whether they are permanent residents or merely visitors. More than 2,000 species have been recorded as fouling organisms, but only about 50 to 100 species are major contributors to the fouling communities of the world. The important species can be divided, for practical considerations, into two groups; organisms with hard (usually calcareous) shells, and organisms without shells. Shelled forms include tubeworms,

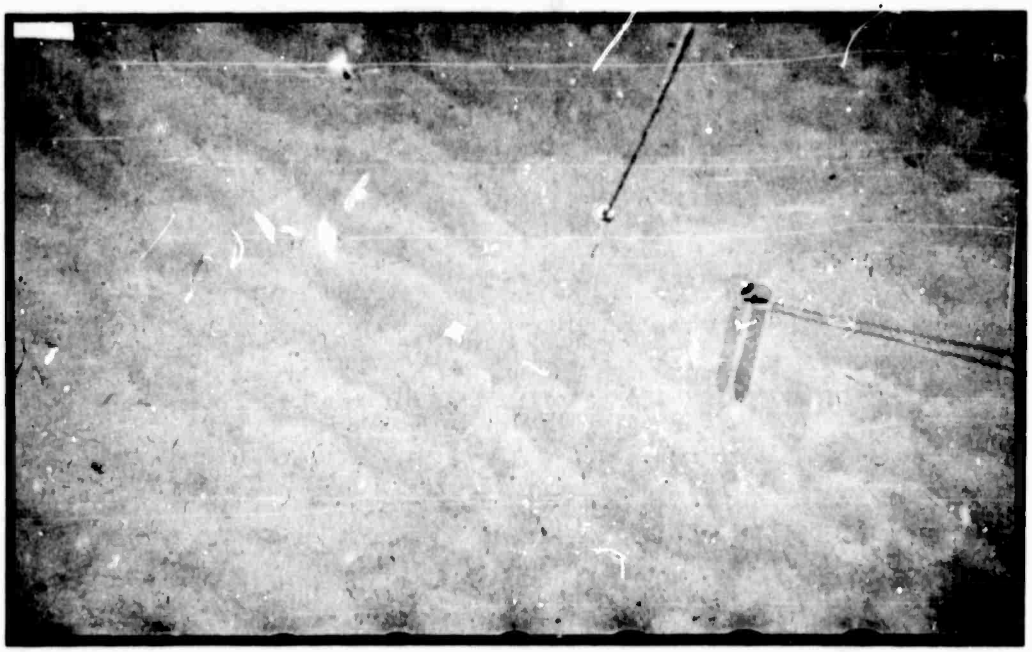


FIGURE 24. BOTTOM PHOTOGRAPHS FROM THE TONGUE OF THE OCEAN. OBSERVE THE STREAMING OF FINER FRAGMENTS SOUTHWARD OF THE LARGE COBBLE ON THE LOWER PORTION OF THE TOP PHOTOGRAPH; ALSO, THE SMOOTHER, MORE PLANATED APPEARANCE OF THE RIPPLE MARKS TO THE RIGHT OF THIS PHOTOGRAPH AS OPPOSED TO THE RIPPLES ON THE BOTTOM PHOTOGRAPH. SCALE APPROXIMATELY 1.34

barnacles, mollusks, and bryozoans. These are particularly important foulers of hulls and sonar domes since their attachment is firm and resists dislocation by water movement. Nonshelled forms such as algae, hydroids, tunicates, and filamentous bryozoans commonly attach to relatively stationary objects such as buoys.

Fouling begins immediately after an object is submerged in water by the initial attachment of microscopic organisms (bacteria and diatoms), which form a slime film, and continues with the successive attachment of larger organisms. The accumulated mass may tear loose when the weight of its resistance to water currents exceeds the holding power of the organisms. The time required for the establishment of a layer of maximum thickness may range from a few months to 3 or more years and is complicated by the different lengths of growing and breeding seasons among the organisms, differences in environmental conditions, and other factors.

Most fouling organisms reproduce by means of the fertilization of eggs by sperms, and the resulting larvae are planktonic (drift with the currents). Duration of the larval stage may vary among species from a few hours to several weeks, but in order to develop into adults the larvae must attach to a suitable surface within this period. Thus, the nearness of a newly exposed surface to an established colony and the speed and direction of the currents play an important part in determining the type of fouling community that will develop. In addition to the length of larval life, the best season for attachment and duration of breeding seasons, which vary among species, also may determine the type and amount of fouling. Thus, objects not exposed during the breeding season of an organism will not be fouled by that organism.

Temperature is the most important factor governing distribution of individual species. Cold may limit colonization by killing the adults. Completion of reproductive cycles may not be possible when warm seasons are too short. Even in tropical waters there may be some temperature limitations in that excessive heat may kill adults or temperature may not fall low enough during the year to permit reproduction of some species. Most organisms also show distinct seasonal variations in reproduction and growth that coincide with temperature changes.

On a local scale the most important environmental factors affecting the composition and development of fouling communities are salinity, pollution, light, and exposure to water movement. Most of the common fouling organisms have a limited tolerance to changes in salinity and live and develop best at normal sea water concentrations of 30 to 35‰. However, a few species can withstand wide salinity variations. Pollution, depending on its type, may either inhibit or facilitate the growth of fouling organisms. Fouling algae and the microscopic plant food supply of many fouling animals are dependent upon light.

The combined influences of the characteristics of fouling organisms and the environmental factors control the time and space distribution of fouling within the ocean itself. Research on fouling leads to several general conclusions concerning its world distribution: 1) Fouling is severest near shore, 2) the greatest accumulation for the shortest time of exposure, more species, and continuous steady attachment is expected to occur in the tropics, 3) fewer species and a short growing season reduce fouling in the arctic, and 4) in

latitudes other than the tropics the fouling community may attain considerable weight and bulk, but requires the longest time of exposure. For example, hydroid, barnacle, and bryozoan communities in the Caribbean may accumulate to a thickness of 2 inches in 40 days, whereas in New England waters mussel accumulations of a similar thickness may require 2 or more years. Generally, the dominant fouler is a different organism in each climatic region.

Sound domes and echo sounder plates may foul rapidly in some areas of the world. This is particularly true of the forward and after ends of the domes where shearing action of the water is least. Accumulation is more rapid on domes of vessels which are less active than those in continuous operation.

The decrease in sound transmission through a dome as a result of fouling is attributed to reflection, scattering, and absorption. Experiments have demonstrated how much fouling can influence transmission of sound through a dome exposed for a relatively short time in waters where fouling is rated severe (Biscayne Bay, Florida). At this location, attenuation of 3 db after 200 days of exposure and 5.5 db after 300 days of exposure was measured, and this attenuation reduced the energy output 25 and 50%, respectively, for a 24 kc sound source. Nonshelled fouling organisms, which may be the dominant foulers in some areas or intermingled with shelled forms in other areas, were discovered to be highly absorptive; that is, they reduce sound energy about 3 db per inch of fouling thickness. The attenuating effects of fouling are greatest during echo ranging, when both the outgoing signal and returning echo must pass through the dome. An additional effect of fouling on listening and ranging is cavitation noise produced by the disruption of water flow around a dome. Nonbiological fouling of stainless steel windows of sound domes frequently occurs as a result of the deposition of bicarbonate ions upon the cathodic stainless steel. This calcareous coating in itself does not significantly influence sound transmission, but it does provide an excellent base for the attachment of fouling organism.

Bioluminescence - Bioluminescence, the emmission of light by living organisms, is a phenomenon of all oceans that has long been observed by mariners. For years the misnomer, phosphorescence, has been used, since it was believed that the light was caused by phosphorus in the water. The phenomenon is produced chiefly by members of the animal kingdom and is rather widely distributed within it. Groups of particular bioluminescent importance are the planktonic flagellates (red tide organisms), coelenterates (jellyfishes), crustaceans (copepods and euphausidds), tunicates (salps), larvae of certain bottom-dwelling animals, squids, and some fishes. Most species are self luminous; that is, they have cells especially adapted for producing light. However, some fishes have luminous bacteria living in open glands on the body which give the same effect as self luminosity.

Bioluminescence may be triggered either by internal or external stimuli. The mechanical disturbances by surface or internal waves; ships, submarines, torpedoes, fish, or whale movements; and upwelling apparently are the most common external stimuli. In an area containing large numbers of luminescent

organisms, bioluminescence might not occur until the onset of one of these stimuli. Bioluminescence is more common in warm tropical waters than in cold northern waters; however, it may be more striking in northern waters during the warm season. Bioluminescent displays have been grouped into 3 basic categories:

(1) Sheet-type - This type usually covers a wide area of the sea surface and is fairly brilliant. It is produced by protozoans, small one-celled plants or animals (especially the dinoflagellates), and possibly by bacteria.

(2) Spark-type - This type takes the form of innumerable points of light appearing intermittently at or near the sea surface. It is produced by small pelagic crustaceans such as copepods, euphausiids, and ostracods.

(3) Globe- or glowing-ball-type - This type consists of numerous globes of light, often densely distributed over a large area of the sea. It is produced by jellyfish, comb jellies, and Pelagic tunicates. While the majority of these organisms are less than one foot in diameter, the volume of illuminated water, which is what usually is seen, may be considerably greater.

Not all plankton are luminescent, but practically all organisms contributing to these displays are plankton. Although these organisms possess some ability to move, they are primarily at the mercy of water currents. Other important environmental factors that affect the distribution and abundance of plankton are temperature, salinity, light, and nutrients.

Sonic Animals - The ability of marine animals to produce sound has long been recognized; however, little research was done on this subject until the importance of sound to naval operations during World War II necessitated an expanded study of all phases of underwater sound. Understanding the basic aspects of the phenomenon requires knowledge of the methods by which the animals produce sound, the species which produce or are capable or producing sound, and the physical characteristics of the sound, such as frequencies and intensities produced. Table 2 lists a few sonic species and their characteristics.

All important sonic marine animals are members of one of 3 groups; crustaceans, fishes, and mammals. Characteristic sounds are produced by animals of each group.

Crustaceans - Crustaceans, particularly snapping shrimp, are one of the most important groups of sonic marine animals. These shrimp bear a general resemblance to the commercial species, but are chiefly distinguishable from them by possession of one large claw which is at least half the size of the body. The characteristic sound of these animals is produced by the snapping of a hard movable finger against the tip of this claw.

Snapping shrimp are distributed in a worldwide belt between approximately 35°N and 35°S (Fig. 25). As may be seen in this figure the principal exception to their distribution is in the Mediterranean Sea and eastern North Atlantic where these shrimp occur in considerable numbers north of 40°N. Actually their geographic distribution appears to be governed by temperature. In general, the 11°C winter isotherm marks the approximate northern and southern limits of their continuous range.

These animals chiefly inhabit bottoms of coral, rock, shell, and often mud or sand when covered by vegetation. Although the largest populations of

Table 2. Sonic characteristics of some important sonic animals.

Name (Common and Scientific)	Frequency range	Principal Frequency	Maximum Pressure (db rel. 1 dyne/cm^2)	Sound description
<u>Invertebrates</u> Snapping shrimp (<u>Alpheus</u>)	1.5 to 45 kc	2 to 20 kc	-29 over shrimp bed; 54 for individual shrimp snap	Persistent background crackling noise over beds, with occasional loud snaps or cracks resembling the sound of burning twigs
<u>Fishes</u> Croakers and drums (<u>Nibes</u> <u>Pseudosciaena</u>)	20 to 1500 cps	50 to 100, and 250 cps	36	Deep drumlike croaks, short toothy rasping
Grunts (<u>Plectorhyschus</u>, <u>Pomadasys</u>)	50 to 4800 cps	200 to 400 and 800 to 1600 cps	35	Scratchy grunting, harsh grating or rasping noises
Gurnards (<u>Chelidonichthys</u>)	40 to 2400 cps	300, 150, 250, and 325 cps	44	Single squawk or series of rapid clucks
Catfishes (<u>Plotosus</u>)	50 to 750 cps	300, 450 cps	No data	Low-pitched grunts
Damselfishes (<u>Abudefduf</u>, <u>Pomacentrus</u>)	20 to 1500 cps	75 to 100 cps	27	Drumming, tapping, repeated snaps; individually feeble, but collectively producing significant background noise
Triggerfishes (<u>Balistes</u>)	50 to 4800 cps	2400 to 4800 cps	26	Metallic scratching; humming and drumming sounds
Jacks (<u>Caranx</u>, <u>Decapterus</u>, <u>Trachurus</u>)	20 to 800 cps	150 to 300, 250, and 350 cps	14	Croaking, clicking, and grating noises

(Continued next page)

Table 2. (Continued)

Name (Common and Scientific)	Frequency range	Principal Frequency	Maximum Pressure (db rel. 1 dyne/cm^2)	Sound description
Mammals Porpoises (*Neomaris*)	from audible range to about 200 kc	7 to 15 kc	70 to 90 (reference level not known)	High-pitched squealing sounds; also whistling, grunting, barking, creaking and clicking; usually a rapid continuous flow of noise

*These data are chiefly the results of laboratory studies involving North American species.

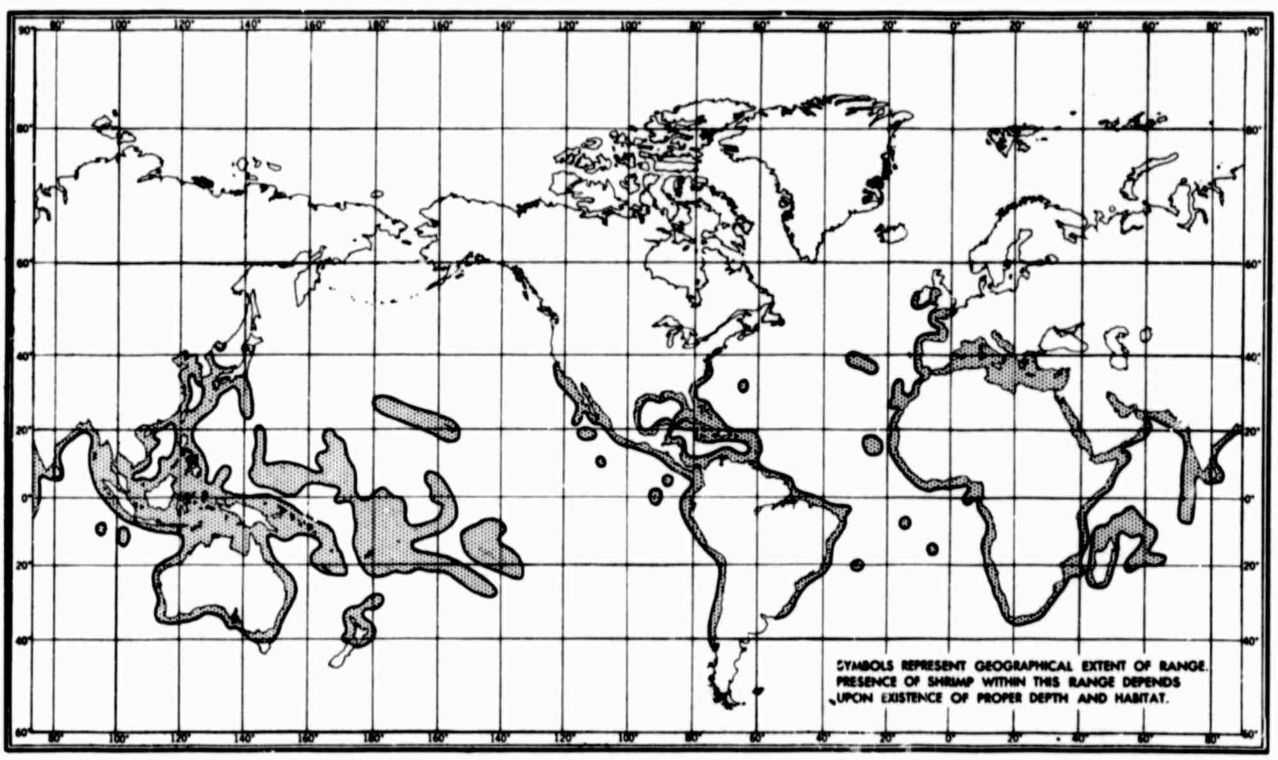


FIGURE 25. DISTRIBUTION OF SNAPPING SHRIMP

shrimp are in depths less than 30 fathoms, shrimp may occur in depths to at least 250 fathoms. However, it is believed that most shrimp noise is produced in water of less than 30 fathoms. Snapping shrimp live together in colonies (beds) and, within their geographic range, tremendous numbers may inhabit localities with a suitable environment.

A shrimp bed is capable of producing an uninterrupted crackle of both sonic and ultrasonic frequencies. This crackle resembles the sound of frying fat or burning underbrush. Although the noise is constant in the vicinity of the bed, a diurnal cycle has been noted, with maximum sound level at sunset. Figure 26 shows some important features of shrimp noise. Among these are: Strongest noise components are between 2 and 20 kc; above 2 kc, shrimp noise completely dominates water noise; and between 10 and 20 kc, shrimp noise is about 26 db above water noise for sea state 2. Acoustic levels drops off rapidly with increasing distance from the bed. Other crustaceans (such as lobsters and crabs) may make minor contributions to background noise.

Fishes - Members of this group are responsible for a variety of sounds, which usually are placed into about 3 categories based upon the method of sound production. The first category includes sounds produced by the air bladder, a membranous sac of atmospheric gases lying in the abdomen. Vibration of this bladder, caused by the movement of muscles within or outside its walls or by general body movement, produces sound, the characteristics of which are determined by the shape and size of the bladder and the manner of excitation.

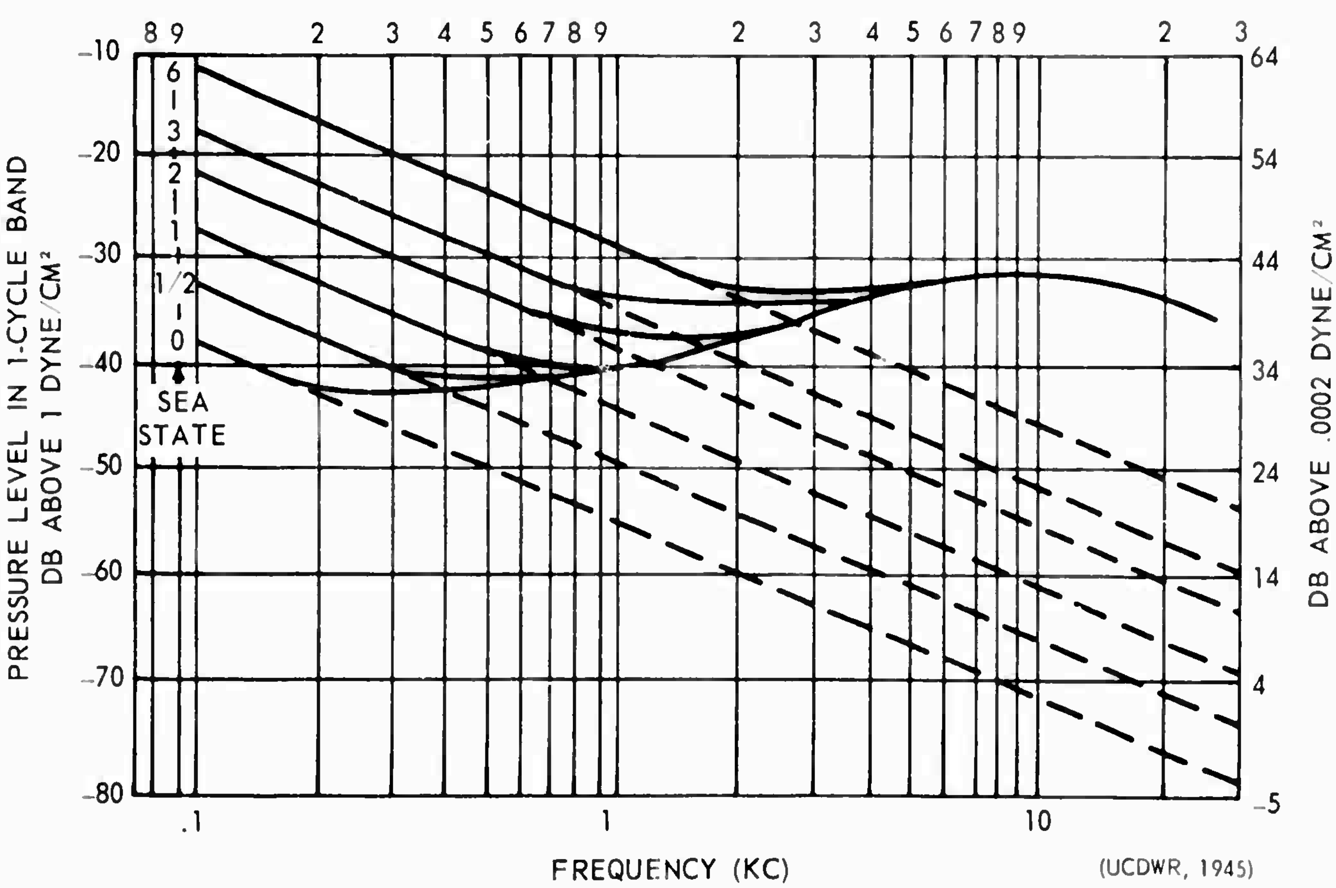


FIGURE 26. AVERAGE AMBIENT NOISE SPECTRA OVER SHRIMP BED

The second category of fish sounds includes those caused by the rubbing together of various back parts of the throat, and is called stridulatory sound. Sounds in these 2 categories generally are considered to be voluntary. A third category includes sounds associated with feeding (biting and chewing) and swimming (collisions between fish or contact with the bottom). These sounds usually are involuntary or incidental to other activities.

Fishes, more than crustaceans, are the source of biological sound in most of the oceans. Their generalized distribution is shown in Figure 27. The families of most important sound producers range principally in warm temperate to tropical waters. In addition, the majority of sonic fishes are inhabitants of coastal waters.

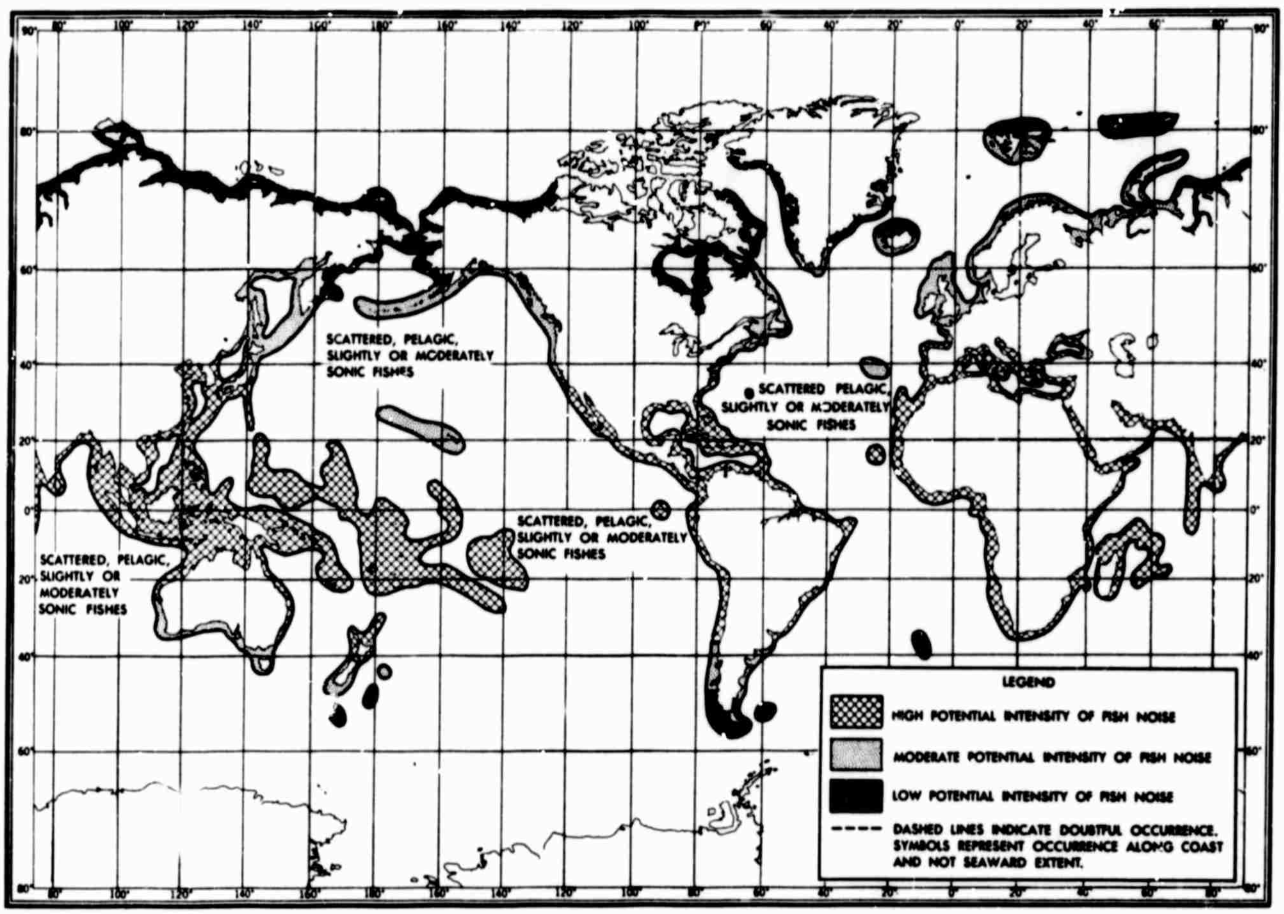


FIGURE 27. ESTIMATED DISTRIBUTION OF FISH NOISE BASED ON THE PRESENCE AND ABUNDANCE OF SONIC SPECIES AND THEIR SONIC ABILITIES

Sonic activity of fishes apparently is not as continuous in intensity as that of snapping shrimp, since in fishes it is associated principally with such periodic activities or conditions as feeding, breeding, and fright. Thus, the most intense sonic activity of many species of fish occurs but once a year, during the breeding season. Also, a daily period or periods of increased sound production, usually at dawn and dusk, often is associated with the feeding habits of many fishes.

Fish sounds range in frequency from about 50 to 8,000 cps. Sounds of air bladder origin have most of their energy concentrated at the lower end of this spectrum, where the more important components are found in the octave band of 75 to 150 cps. Stridulatory sounds characteristically are concentrated at the higher end of this spectrum.

Marine mammals - In addition to returning echoes from sound ranging and depth finding gear, marine mammals also produce sound. In a manner similar to that of terrestrial mammals, marine mammals may produce sound vocally. In whales and porpoises characteristic sounds are made by air forced through soft-walled nasal sacs, often without emission through the blowhole. Sounds probably originating from this source have been described as echo-ranging pings, squealing, and long drawn out moaning similar to a foghorn blast.

Seals, sea lions, and similar animals possess vocal cords and produce their barks and whistles by means of them. Hisses and snorts are produced by the forcible expelling of air through the mouth and nostrils.

Other sounds of probable internal origin are produced by snapping together of the animal's jaws or by stridulation of the teeth or the plates of baleen. Sounds possible originating from these sources have been related principally to the whales and propoises and are described as clicking and snapping.

Sounds of external origin are produced chiefly by the action of the flukes and flippers during swimming. These sounds, especially those from the larger whales, may resemble propeller or engine noises.

The toothed whales, which include the sperm whale, killer whale (actually a porpoise), and porpoise are considerably more noisy than the whalebone (baleen) whales such as the blue, humpback, and gray. The difference is related to the internal formation of the nasal region in the two groups. However, despite their relative vocal silence, the whalebone whales are important producers of swimming sounds.

Mammals, especially the whales, may contribute to biological sound in all waters. Sonic crustaceans are limited to the bottom in shallow coastal waters, and the most important sonic fishes frequent the water column in this same general zone. Mammal sounds include a much greater range of frequencies than that of either crustaceans or fishes. They have been recorded as low as 70 cps and possibly lower (whale sounds) and as high as 196 kc (porpoise sounds), although the principal frequencies are in the audible range.

Biological Scattering - Some marine organisms play an important role in underwater acoustics, primarily because of their sound scattering properties.

The sound-scattering potential of fishes has been recognized since at least the early 1930's, when an echo sounder was first used to locate schools of fish. Sound scattering by other marine organisms was discovered in more recent years.

Evidence of a deep scattering layer (DSL) first was observed and reported by the Echo Ranging Group of the University of California Division of War Research during World War II. This phenomenon, as originally noted and since typically recorded on standard echo sounders using 10- to 12-kc sound, appears on recording paper as a relatively narrow trace displaying diurnal vertical movement. Soon after its discovery, a relationship between this movement and similar movements of various species of plankton and fish was noted. This theory started many attempts to find, by several means such as net hauls, trawling, and underwater photography, the cause of the scattering as well as to undertake extensive research on the sound-scattering properties of marine plankton and fishes. Results of this research generally have been promising, but inconclusive. However, on the basis of research to the present, the shrimplike euphausiids and certain deep-living fishes appear to be the most important contributors to sound scattering in the DSL.

The DSL has been recorded in all oceans, although conflicting reports have been made concerning its presence in antarctic waters. A typical appearance on a standard echo sounder (Fig. 28) seems to be a simple trace lying in a depth range of approximately 150 to 250 fathoms during the day, ascending to near the surface at about sunset, and descending to depth at about sunrise. This pattern is misleading, as on high resolution sounding machines and during specially devised tests a more complicated structure of the DSL has been recorded. Instead of a simple trace, the record has shown many horizontal traces and patches. These suggest the occurrence of scattering organisms in numerous layers that are more or less separate from each other and that live in masses or schools. Maximum daytime depths of 400 fathoms and the occasional failure of one or more of these sublayers to move vertically also have been observed.

Evidence of the apparent negative reaction of the organisms to light, demonstrated by the descent of the layer at sunrise, is strengthened by the behavior of this layer under other circumstances. For example, clear indication of the trace and depth of the layer during daylight hours varies with weather and latitude. In any one area the layer usually is more prominent when the sky is clear than when overcast. Also, under polar conditions of permanent daylight the layer seldom is noted or at best is weakly defined. However, when normal day and night conditions exist, the layer again is noticeable.

The DSL when detected by echo sounders is rarely acoustically opaque; that is, the bottom, when in range, usually is recorded beneath it. However, sound will be attenuated when directed horizontally through a large concentration of scatterers.

Marine organisms with scattering properties are also permanent inhabitants of the surface layers in deep water; others are abundant in shallow coastal waters. Concentrations of the latter group, dense enough to return traces observable on depth recorders, often are referred to as the shallow scattering layer, although this group does not undergo diurnal vertical migration that is associated with the DSL.

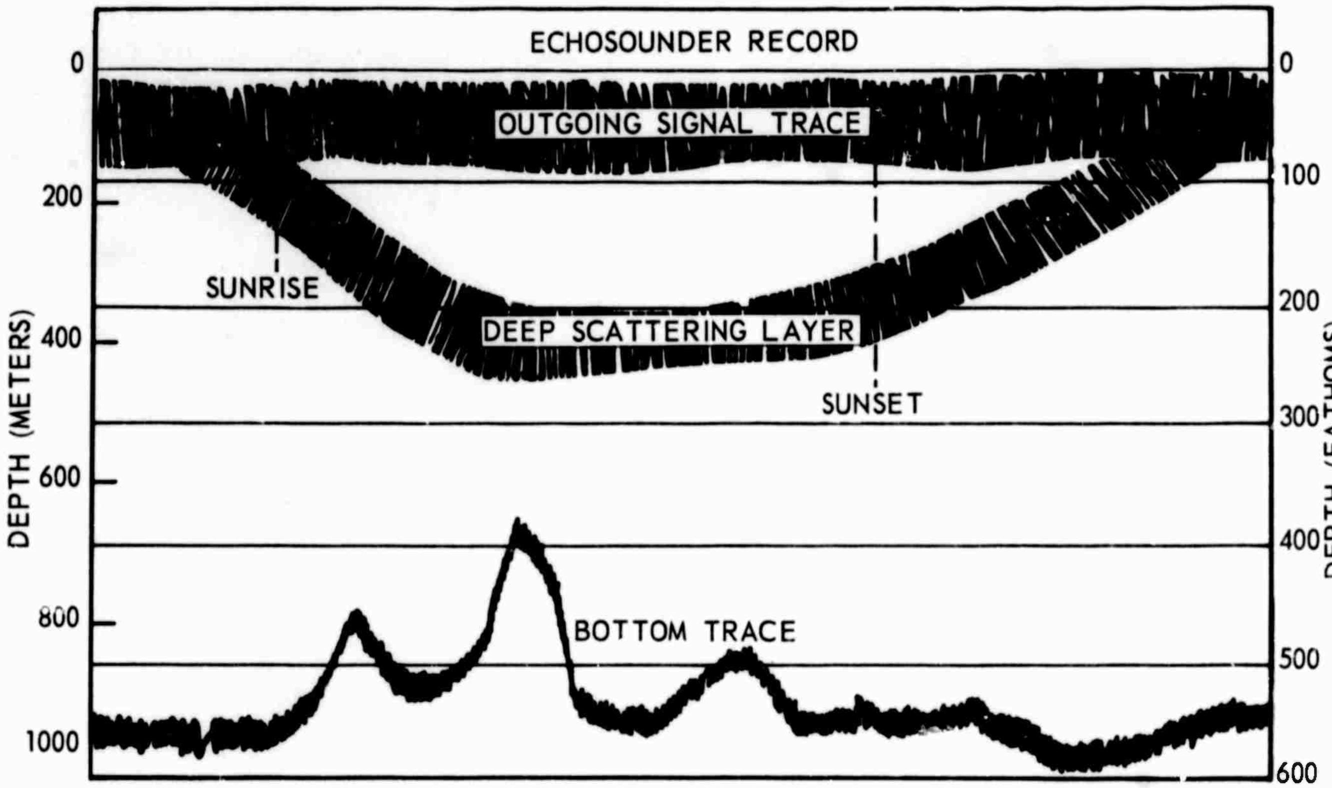


FIGURE 28. REPRESENTATION OF THE DEEP SCATTERING LAYER (DSL)

False Targets - Whales, schools of fishes, herds of porpoises, detached masses of seaweed, and dense concentrations of plankton have been implicated or suspected in reports of false sonar targets. The size of the individual or group and its sound reflecting ability are important factors in considering any species as a potential false target. For example, small fish swimming in a dense school may appear acoustically as a single unit.

Knowledge of the nature and habits of animals allows prediction of their geographic and seasonal distribution. This is especially true of most whales and many fishes that exhibit definite migratory patterns. Whales are found along the coasts and in the open sea of all oceans. The majority of the whalebone (baleen) whales undertake regular seasonal migrations during which they feed principally in the plankton-rich, cold polar waters in summer and breed in the warm waters toward the Equator in winter. However, the extent of seasonal movement varies considerably among these whales. For example, the bowhead whale which inhabits the arctic region migrates very little if at all. Compared to this the California gray whale, which makes the longest migration of all whales, travels 5,000 to 7,000 miles from its feeding grounds in the Bering Sea to its breeding grounds along lower California. Migrations are not undertaken en masse by the whales in a region. In fact along some routes, notably that of the gray whale, late arrivals to the breeding grounds often pass early departing individuals traveling to the feeding grounds.

The sperm whale (a toothed whale) is distributed over a large area of the world's oceans. However, since its principal food (giant squid) is

especially abundant in tropical and temperate waters, this whale is found mainly within these climatic regions. Within this range sperm whales display some poleward movement with spring warming trends. Other toothed whales or porpoises are abundant throughout the world. The killer whale, although especially plentiful in polar regions, is worldwide in distribution. This vicious attacker of almost any marine animal, including even the large whales, often travels in packs of as many as 40 individuals. Porpoises chiefly frequent nearshore waters but may travel at least several hundred miles from land. Schools of thousands of individuals have been reported; many of these reports are from the eastern Pacific off the American coasts. Considerable variation in size, social relationships, and swimming characteristics exists among the various whales and porpoises. Table 3 lists many of the characteristics of these animals.

TABLE 3 DISTRIBUTION AND CHARACTERISTICS OF CERTAIN MARINE MAMMALS AND FISHES

NAME	DISTRIBUTION	SIZE (FEET)	GREGARIOUSNESS	SWIMMING SPEED (KNOTS)	DIVING AND SURFACING *
WHALEBONE WHALES					
BLUE	WORLDWIDE, USUALLY IN OPEN WATERS	75 (TO 100 IN ANTARCTIC)	SINGLY OR IN PAIRS, LESS FREQUENTLY IN GROUPS TO 10 ANIMALS	10-15, NORMAL 14-15, FOR 5 HOURS TO 20, FOR 10 MINUTES TO 20, SPURTS	AFTER SURFACING FROM DEEP DIVE, GENERALLY MAKES DOZEN OR MORE SHALLOW DIVES 13-15 SECONDS DURATION; DEEP DIVE LASTS 10-20 MINUTES, OBSERVED MAXIMUM OF 50 MINUTES **
CALIFORNIA GRAY	PACIFIC OCEAN, NORTH OF 20°N ***	35-45	USUALLY PAIRS OR GROUP OF 5, OFTEN 3 OR 6 ANIMALS, RARELY 10 OR MORE ANIMALS	4-5, AVERAGE 7-8, MAXIMUM	BLOW OF 5 SECONDS FOLLOWED BY SHORT DIVE OF 10-15 SECONDS, THIS USUALLY REPEATED 5 TIMES BEFORE LONG DIVE OF 3-7 MINUTES; MAXIMUM DIVE 10 MINUTES AND/OR 500-FOOT DEPTH
FIN	WORLDWIDE ***	60-70 (TO 85 IN ANTARCTIC)	SINGLY OR PAIRS, OFTEN GROUPS OF AS MANY AS 300 ANIMALS	10-15, NORMAL 16-20, MAXIMUM	TAKES 3-15 BREATHES BEFORE DIVING FOR 8-15 MINUTES, MAXIMUM DIVE OF 30 MINUTES
HUMPBACK	WORLDWIDE ***	40-50 (TO 65 MAX.)	SMALL GROUPS OF 4-8 ANIMALS, GROUPS TO 20 ANIMALS	4.5, AVERAGE 3.5-6.5, RANGE	ONE TO 18 SHALLOW DIVES BEFORE DEEP DIVE, DEEP DIVE DURATION 15-20 MINUTES
RIGHT	WORLDWIDE, MAINLY TEMPERATE WATERS ***	50-60	SINGLY AND SMALL GROUPS	5, SURFACE MAXIMUM 6-8, SOUNDING	DURATION OF DIVE 15-20 MINUTES, MAXIMUM OF 60 MINUTES
SEI	WORLDWIDE	40-55	GROUPS (?)	30, SPURTS OF LESS THAN ONE-HALF MILE	NO DATA
TOOTHED WHALES					
SPERM	WORLDWIDE, CHIEFLY BETWEEN 40° NORTH AND SOUTH	60, MALES (MAX.) 30-40, FEMALES	GROUPS TO 400 ANIMALS, OLD BULLS SINGLY OR IN SMALL GROUPS	3-4, AVERAGE 10-12, MAXIMUM 20, SPURTS	BLOWS 30-35 TIMES (TO 60-70 TIMES) FOR 10-11 MINUTES; DIVES TO 500 FATHOMS FOR 40-50 MINUTES, MAXIMUM DIVE 656 FATHOMS FOR 75 MINUTES
PORPOISES					
KILLER WHALE	WORLDWIDE, MAINLY IN WATERS FREQUENTED BY OTHER MARINE MAMMALS	20-30, MALES 15-20, FEMALES	GROUPS OF 5-40	15-20, NORMAL 30, MAXIMUM	CHIEFLY SURFACE OR NEAR SURFACE WATERS
PILOT WHALE (BLACKFISH)	WORLDWIDE	20-30	LARGE GROUPS TO OVER 1,000 ANIMALS	NO DATA	NO DATA
OTHER PORPOISES	WORLDWIDE, MAINLY IN COASTAL AREAS	5-10	GROUPS TO OVER 1,000 ANIMALS	15-30, NORMAL 40, MAXIMUM	USUALLY SURFACE EVERY 30-45 SECONDS, MAY SUBMERGE FOR 3-7 MINUTES
SHARKS					
BASKING SHARK	TEMPERATE WATERS	40-50	SINGLY OR PAIRS, GROUPS OF [illegible]-100 ANIMALS	SLOW	USUALLY ON SURFACE, MAY SUBMERGE TO 200 FATHOMS
WHALE SHARK	WARMER WATERS OF MAJOR OCEANS	40-60	SINGLY OR GROUPS	SLOW	USUALLY ON SURFACE

* THE DIVING AND SURFACING CHARACTERISTICS WILL VARY IF THE WHALE IS DISTURBED OR FRIGHTENED
** WHALEBONE WHALES RARELY SUBMERGE DEEPER THAN 50 FATHOMS
*** THOUGH ALL WHALES MAY ENTER COASTAL WATERS OCCASIONALLY, THESE SPECIES DO SO MORE FREQUENTLY

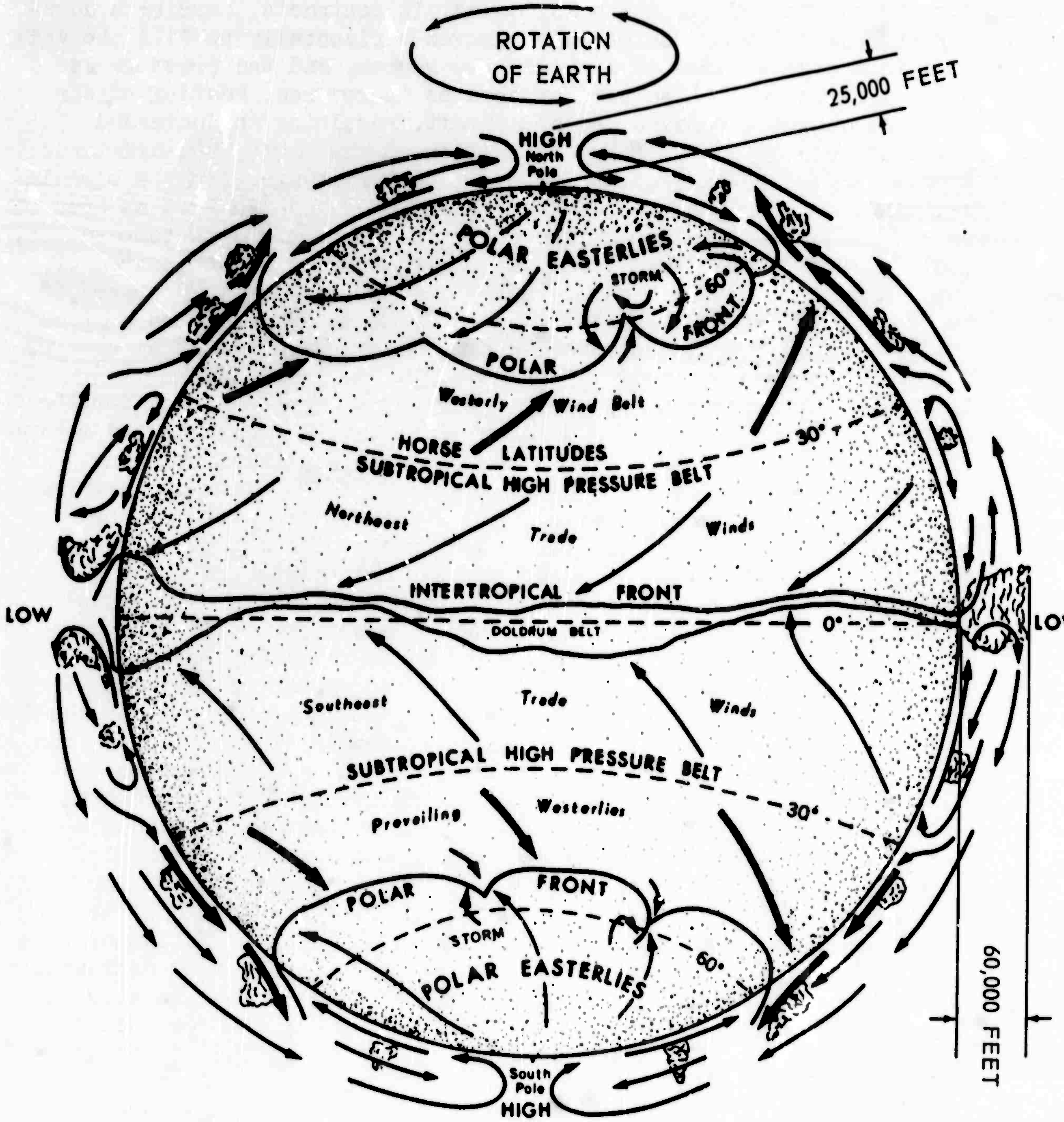


FIGURE 29. GENERAL WIND PATTERNS OF THE EARTH

some distance above the surface of the earth, the wind tends to blow along isobars (lines of equal atmospheric pressure). Near the surface of the earth, friction tends to divert the wind from the isobars toward the center of low pressure. At sea, where friction is less than on land, the wind follows the isobars more closely.

A thin belt of low pressure near the Equator occupies a position approximately midway between high pressure belts at about latitude 30° to 35° on each side of the Equator (Fig. 29). Except for slight diurnal changes, the atmospheric pressure along the equatorial low is almost uniform, and the wind is practically nonexistent. The light breezes that do blow are variable in direction. Hot, sultry days are common. The sky is often overcast, and showers and thundershowers are relatively frequent. This area is called the doldrums. The eastern part of the belt in both the Atlantic and Pacific is wider than the western part; however, both the position and extent of the belt vary somewhat with the season. During February and March, it lies immediately to the north of the Equator and is so narrow that it may be considered virtually nonexistant. In July and August, the belt is centered on about latitude 7°N and is several degrees in width, even at the narrowest point.

The trade winds blow from the belts of high pressure toward the equatorial belt of low pressure. Because of the rotation of the earth, the moving air is deflected toward the west. Therefore, the trade winds in the Northern Hemisphere are from the northeast and are called the northeast trades, while those in the Southern Hemisphere are from the southeast and are called the southeast trades. Over the eastern part of both the Atlantic and Pacific, these winds extend considerably farther from the Equator, and their original direction is more nearly along the meridians than in the western part of each ocean.

The trade winds are among the most constant of winds. They sometimes blow for days or even weeks with little change of direction or speed. At times they weaken or shift direction, and there are regions where the general pattern is disrupted. A notable example is the island groups of the South Pacific where they are practically nonexistent during January and February. Their highest development is attained in the South Atlantic and the South Indian Oceans. Everywhere they are fresher during the winter than during the summer.

In July and August, when the belt of equatorial low pressure moves to a position some distance north of the Equator, the southeast trades blow across the Equator into the Northern Hemisphere where the earth's rotation diverts them toward the right, causing them to be southerly and southwesterly winds. The southwest monsoons of the African and Central American coasts have their origin partly in such diverted winds. Cyclonic storms generally do not enter the regions of the trade winds, although hurricanes and typhoons may originate within these areas.

Along the poleward side of each trade-wind belt, corresponding approximately with the belt of high pressure in each hemisphere, is another region with weak pressure gradients and correspondingly light, variable winds. These are called the horse latitudes. The weather is generally clear and fresh, unlike that in the doldrums, and periods of stagnation are less persistent, being of a more intermittent nature. The difference is due primarily to rising

currents of warm air in the equatorial low carrying large amounts of moisture which condenses as the air cools at higher levels, while in the horse latitudes the air is apparently descending and becoming less humid as it is warmed at lower heights.

On the poleward side of the high pressure belt in each hemisphere, the atmospheric pressure again diminishes. The currents of air set in motion along these gradients toward the poles are diverted by the earth's rotation toward the east, becoming southwesterly winds in the Northern Hemisphere and northwesterly in the Southern Hemisphere. These two wind systems are known as the prevailing westerlies of the temperate zones.

In the Northern Hemisphere this relatively simple pattern is distorted considerably by secondary wind circulations, due primarily to the presence of large land masses. In the North Atlantic, between latitudes 40° and 50°N, winds blow from some direction between south and northwest 74 % of the time and are somewhat more persistent and stronger in the winter than in summer. In the Southern Hemisphere the westerlies blow throughout the year with a steadiness approaching that of the tradewinds. The speed, though variable, is generally between 17 and 27 knots. Latitudes 40° to 50°S (or 55°S), where these boisterous winds occur, are called the roaring forties. The greater speed and persistence of the westerlies below the Equator is due to the comparatively landlessness of the Southern Hemisphere. Also, the average yearly atmospheric pressure diminishes much more rapidly on the poleward side of the high pressure belt, and has fewer irregularities due to continental interference than in the Northern Hemisphere.

Because of the low temperature near the geographical poles of the earth, the pressure tends to remain higher than in surrounding regions. Consequently, the winds blow outward from the poles, and are deflected westward by the rotation of the earth, to become northeasterlies in the arctic, and southeasterlies in the antarctic. Where these meet the prevailing westerlies, the winds are variable.

In the arctic, the general circulation is greatly modified by surrounding land masses. Winds over the Arctic Ocean are somewhat variable, and strong surface winds are rarely encountered. In the antarctic, on the other hand, a high central land mass is surrounded by water, a condition which augments, rather than diminishes the general circulation. The high pressure, although weaker than in some areas, is stronger than in the arctic, and of great persistence near the South Pole. The upper air descends over the continent, where it becomes intensely cold. As it moves outward and downward toward the sea, it is deflected toward the west by the earth's rotation. The winds remain strong throughout the year, frequently attaining hurricane force, and sometimes reaching speeds of 100 knots at the surface.

The general circulation of the atmosphere as described above is greatly modified by various conditions. The high pressure in the horse latitudes is not uniformly distributed around the belts, but tends to be accentuated at several points, as shown in Figures 13 and 14. These semipermanent highs remain at about the same places with great persistence.

Semipermanent lows also occur in various places, the most prominent ones being west of Iceland and over the Aleutians (winter only) in the Northern Hemisphere, and at the Ross Sea and Weddell Sea in the antarctic. The areas

occupied by these semipermanent lows are sometimes called the graveyards of the lows, since many lows move directly into these areas and lose their identity as they merge with and reinforce the semipermanent lows. The low pressure in these areas is maintained largely by the migratory lows which stall there, but partly by the sharp temperature difference between polar regions and warmer ocean regions.

Another modifying influence is land, which undergoes greater temperature changes than does the sea. During the summer, a continent is warmer than its adjacent oceans; therefore, low pressures tend to prevail over the land. If a belt of high pressure encounters such a continent, its pattern is distorted or interrupted, and the belt of low pressure is intensified. The winds associated with belts of high and low pressure are distorted accordingly. In winter, the opposite effect takes place, belts of high pressure being intensified over land and those of low pressure being interrupted.

The most striking example of a wind system produced by the alternate heating and cooling of a land mass is the monsoons of the China Sea and Indian Ocean. In the summer, low pressure prevails over the warm continent of Asia and high pressure over the adjacent sea. Between these two systems the wind blows in a nearly steady direction. The lower portion of the pattern is in the Southern Hemisphere, extending to about 10°S. Here the rotation of the earth causes a deflection to the left, resulting in southeasterly winds. As they cross the Equator, the deflection is in the opposite direction, causing them to curve toward the right, becoming southwesterly winds. In the winter the position of high and low pressure areas are interchanged, and the direction of flow is reversed.

In the China Sea the summer monsoon blows from the southwest, usually from May to September. The strong winds are accompanied by heavy squalls and thunderstorms, the rainfall being much heavier than during the winter monsoon. As the season advances, squalls and rain become less frequent. In some places the wind becomes a light breeze which is unsteady in direction, or stops altogether, while in other places it continues almost undiminished with changes in direction or calms being infrequent. The winter monsoon blows from the northeast, usually from October to April, with a steadiness similar to that of the trade winds, often attaining the speed of a moderate gale (28-33 knots). Skies are generally clear during this season, and there is relatively little rain.

The general circulation of the earth is further modified by winds of cyclonic origin and various local winds. As air masses move within the general circulation, they travel from their source regions and invade other areas dominated by air having different characteristics. There is little tendency for adjacent air masses to mix. Instead, they are separated by a thin zone in which air mass characteristics exhibit such sharp gradients as to appear as discontinuities called frontal surfaces. The intersection of a frontal surface and a horizontal plane is called a front. Before the formation of frontal waves, the isobars (lines of equal atmospheric pressures) tend to run parallel to the fronts. As a wave is formed, the pattern is distorted somewhat and waves tend to travel in the direction of the general circulation, which in the temperate latitudes is usually in a general easterly and slightly poleward direction. Along the leading edge of the wave,

warmer air is replacing colder air. This is called the warm front. The trailing edge is the cold front, where colder air is replacing warmer air.

The warm air, being less dense, tends to ride up over the colder air it is replacing, causing the warm front to be tilted in the direction of motion. The slope is gentle, varying between 1:100 and 1:300. Because of the replacement of cold, dense air with warm, light air, the pressure decreases. Since the slope is gentle, the upper part of a warm frontal surface may be hundreds of miles ahead of the surface portion. The decreasing pressure, indicated by a falling barometer, is often an indication of the approach of such a wave. In a slow-moving, well-developed wave, the barometer may begin to fall several days before the wave arrives. Thus, the amount and nature of the change of atmospheric pressure between observations, called pressure tendency, is of assistance in predicting the approach of such a system.

The advancing cold air, at the trailing edge being more dense, tends to cut under the warmer air at the cold front, lifting it to greater heights. The slope here is in the opposite direction, at a rate of about 1:25 to 1:100, and is steeper than the warm front; therefore, after a cold front has passed, the pressure increases.

The approach of a well-developed warm front is usually heralded not only by falling pressure, but also by a more or less regular sequence of clouds. First, cirrus appear. These give way successively to cirrostratus, altostratus, altocumulus, and nimbostratus. Brief showers may precede the steady rain accompanying the nimbostratus.

As the warm front passes, the temperature rises, the wind shifts to the right (in the Northern Hemisphere), and the steady rain stops. Drizzle may fall from low-lying stratus clouds for some time after the wind shifts. During passage of the warm sector between the warm front and the cold front, there is little change in temperature or pressure. However, if the wave is still growing and the low deepening, the pressure might decrease slowly. In the warm sector the skies are generally clear or partly cloudy, with cumulus or stratocumulus clouds most frequent. The warm air is usually moist, and haze or fog may often be present.

As a faster-moving, steeper cold front passes, the wind shifts abruptly to the right (in the Northern Hemisphere), the temperature falls rapidly, and there are often brief and violent showers, frequently accompanied by thunder and lightning. Clouds are usually of the convective type. A cold front usually coincides with a well-defined wind-shift line (a line along which the wind shifts abruptly from southerly or southwesterly to northerly or northwesterly in the Northern Hemisphere and from northerly to northwesterly to southerly or southwesterly in the Southern Hemisphere). At sea a series of brief showers accompanied by strong, shifting winds may occur along or some distance (up to 200 miles) ahead of a cold front. These are called squalls. Because of its greater speed and steeper slope, which may approach or even exceed the vertical near the earth's surface (due to friction), a cold front and its associated weather passes more quickly than a warm front. After a cold front passes, the pressure rises, often quite rapidly, the visibility usually improves, and the clouds diminish.

When a faster-moving cold front overtakes a warm front an occluded front at the surface is formed. When the two parts of the cold air mass meet, the

warmer portion tends to rise above the colder part. The warm air continues to rise until the entire system dissipates. As the warmer air is replaced by colder air, the pressure gradually rises.

Although each low follows generally the pattern given above, no two are ever exactly alike. Other centers of low pressure and high pressure and the air masses associated with them, even though they may be 1,000 miles or more away, influence the formation and motion of the individual low centers and their accompanying weather.

Because of the rotation of the earth, the circulation tends to be counterclockwise (cyclonic) around areas of low pressure in the Northern Hemisphere and clockwise (anticyclonic) around areas of high pressure, the speed being proportional to the spacing of isobars.

Since an anticyclone area is a region of outflowing winds, air is drawn into it from aloft. Descending air is warmed, and as air becomes warmer, its capacity for holding uncondensed moisture increases. Therefore, clouds tend to dissipate. Clear skies are characteristic of an anticyclone, although scattered clouds and showers are sometimes encountered.

In contrast, a cyclonic area is one of converging winds. The resulting upward movement of air results in cooling, a condition favorable to the formation of clouds and precipitation. More or less continuous rain and generally stormy weather are usual with a cyclone.

Between the two belts of high pressure associated with the horse latitudes, cyclones form only occasionally, generally in certain seasons, and always in certain areas at sea. These tropical cyclones are usually quite violent, being known under various names according to their location.

In the areas of the prevailing westerlies, cyclones are a common occurrence, the cyclonic and anticyclonic circulation being a predominant feature of temperate latitudes. These are sometimes called extratropical cyclones to distinguish them from the more violent tropical cyclones. Although most of them are formed at sea, their formation over land is not unusual. As a general rule, they decrease in intensity when they encounter land, and increase when they move from the land to a water area. In their early stages, cyclones are elongated, but as their life cycle proceeds, they become more nearly circular.

In addition to the winds of the general circulation and those associated with cyclones and anticyclones, there are numerous local winds which influence the weather in various places.

ICE IN THE SEA

The perpetually-frozen Arctic Ocean and the solid sheet of ice beneath which Antarctica is buried offer evidence that the earth has not yet completely emerged from its most recent Ice Age. Each winter this polar ice increases and spreads toward more temperate latitudes, and each summer it contracts again as part of the ice melts.

As it cools, water contracts until the temperature of maximum density (4°C) is reached. Further cooling results in expansion. The addition of salt lowers both the temperature of maximum density and, to a lesser extent, that of freezing. At a salinity of 35 ‰, the approximate average for the oceans, the freezing point is -1.9°C.

Ice forms first at the surface. As it forms, most of the dissolved solids remain in the water beneath the ice, increasing the density of the water there. This lowers the freezing point, thus tending to retard the freezing process. It is further retarded by the fact that ice is a poor conductor of heat and therefore serves as an insulator to protect the water from colder air above.

In shoal water and streams, particularly where motion is sufficient to cause thorough mixing, the freezing temperature may extend from the surface to the bottom. When this occurs, ice crystals may form at any depth. Because of their decreased density, they tend to rise to the surface, unless they form at the bottom and attach themselves there. Ice may also be formed by the compacting of fallen snow, or by the freezing of a mixture of snow and sea water.

The first indication of sea ice is a greasy or oily appearance of the surface, with a peculiar gray or leaden tint. The small individual particles of ice, called spicules, then become visible. As the number increases, the mixture of water and ice is soupy or mushy, having about the consistency of wet snow and is called slush. As the individual particles freeze together, a thin layer of highly plastic ice forms. This bends easily and moves up and down with waves. A layer of 2 inches of fresh-water ice is brittle but strong enough to support the weight of a heavy man. In contrast, the same thickness of newly-formed sea ice will support not more than about 10 % of this weight, although its strength varies with the temperature at which it is formed, very cold ice supporting a greater weight than warmer ice. When snow falls into sea water which is near its freezing point but colder than the melting point of snow, it does not melt, but floats on the surface, drifting with the wind into beds which may become several feet thick. If the temperature drops below the freezing point of the sea water the mixture of snow and water freezes quickly into soft ice similar to that formed when snow is not present. As it ages, sea ice becomes harder and more brittle.

Sea ice is exposed to several forces, including currents, wave motion, tides, wind, and temperature differences. In its early stages, its plasticity permits it to conform readily to virtually any shape required by the forces acting upon it. As it becomes older, thicker, and more brittle, exposed sea ice cracks and breaks under the strains. Under the influence of wind and current, the broken pieces may shift position relative to pieces around them.

When ice is formed in the presence of considerable wave motion, circular pancakes several feet in diameter are formed, rather than a single large sheet. Wave motion may cause the pancakes to break into smaller pieces. With continued freezing, individual pieces unite into floes, and floes into ice fields which extend over many miles.

When one floe encounters another, or the shore, the individual pieces may be forced closer together into a thickly-compacted mass. If the force is sufficient, and the ice is sufficiently plastic, bending takes place, or tenting if the contacting edges of individual cakes force each other to rise above their surroundings. More frequently, however, rafting occurs, as one cake overrides another. Sea ice having any readily observed roughness of the surface is called pressure ice. A line of ice piled haphazardly along the edge of two floes which have collided is called a pressure ridge. Pressure ice with numerous mounds or hillocks which have become somewhat rounded and

smooth by weathering or the accumulation of snow is called hummocked ice.

A large mass of sea ice, consisting of various floes, pressure ridges, and openings, is called a pack. In the arctic the main pack extends over the entire Arctic Ocean and for a varying distance outward from it, the limits receding considerably during summer. Each year a large portion of the ice from the Arctic Ocean moves outward between Greenland and Norway, into the North Atlantic, and is replaced by new ice. Relatively little of the pack ice is more than 10 years old. The ice pole, the approximate center of the arctic pack, is at 83.5°N, 160°W (north of western Alaska and about 390 miles from the North Pole). In the antarctic the pack exists as a relatively narrow strip between the continent of Antarctica and the notoriously stormy sea which hasten the pack's destruction.

The alternate melting and refreezing of the surface of the pack, producing weathered ice, combined with the various motions to which the pack is subjected, result in widely varying conditions within the pack itself. The extent to which it can be penetrated by a ship varies from place to place and with changing weather conditions. In some areas the limit of navigable water is abrupt and complete, as at the edge of shelf ice.

Although surface currents have some effect upon the drift of pack ice, the principal factor is wind. Due to Coriolis force, ice does not drift in the direction of the wind, but about 30° from this direction. In the Northern Hemisphere this drift is to the right of the direction toward which the wind blows, and in the Southern Hemisphere it is toward the left. Since the surface wind is deflected about twice this amount from the direction of the pressure gradient, the total deflection of the ice is about 90° from the pressure gradient. The rate of drift is about 1 to 7 % of the wind speed, depending upon the roughness of the surface and the concentration of the ice.

The underside of pack ice reflects approximately its surface topography. If the topography is flat or the hummocks weathered, then the bottom will be smooth. If the surface is ridged, hummocked, and rough, then the bottom will have counterparts but will be much smoother and more rounded compared to the surface. Protuberances below the water may reach depths of 40 to 60 or more feet in summer.

When a glacier flows into the sea, the buoyant force of the water breaks off pieces from time to time, and these float away as icebergs. An iceberg seldom melts uniformly because of lack of uniformity in the ice itself, differences in the temperature above and below the water line, exposure of one side to the sun, strains, cracks, mechanical erosion, etc. The inclusion of rocks, silt, and other foreign matter further accentuates the differences. As a result, changes in equilibrium take place, which may cause the berg to tilt or capsize. Parts of it may break off or calve, forming separate, smaller bergs. A small berg about the size of a house is called a bergy bit, and one still smaller but large enough to inflict serious damage to a vessel is called a growler because of the noise it sometimes makes as it bobs up and down in the sea. Bergy bits and growlers are usually pieces calved from icebergs, but they may be formed by consolidation of sea ice or by the melting of an iceberg.

Icebergs which extend a considerable distance below the surface and have a relatively small sail area are influenced more by surface currents than by wind. However, if a strong wind blows for a number of hours in a steady direction, the drift of icebergs will be materially affected. In this instance the effect is two-fold. The wind acts directly against the iceberg, and also generates a surface current in about the same direction. Because of inertia, an iceberg may continue to move from the influence of wind for some time after the wind stops or changes direction.

Ice formation in the ocean, in common with other meteorological and oceanographic phenomena, varies considerably from year to year, and wide deviations from average conditions are not unusual. Figure 30 gives information on the average limits of ice in the Northern Hemisphere.

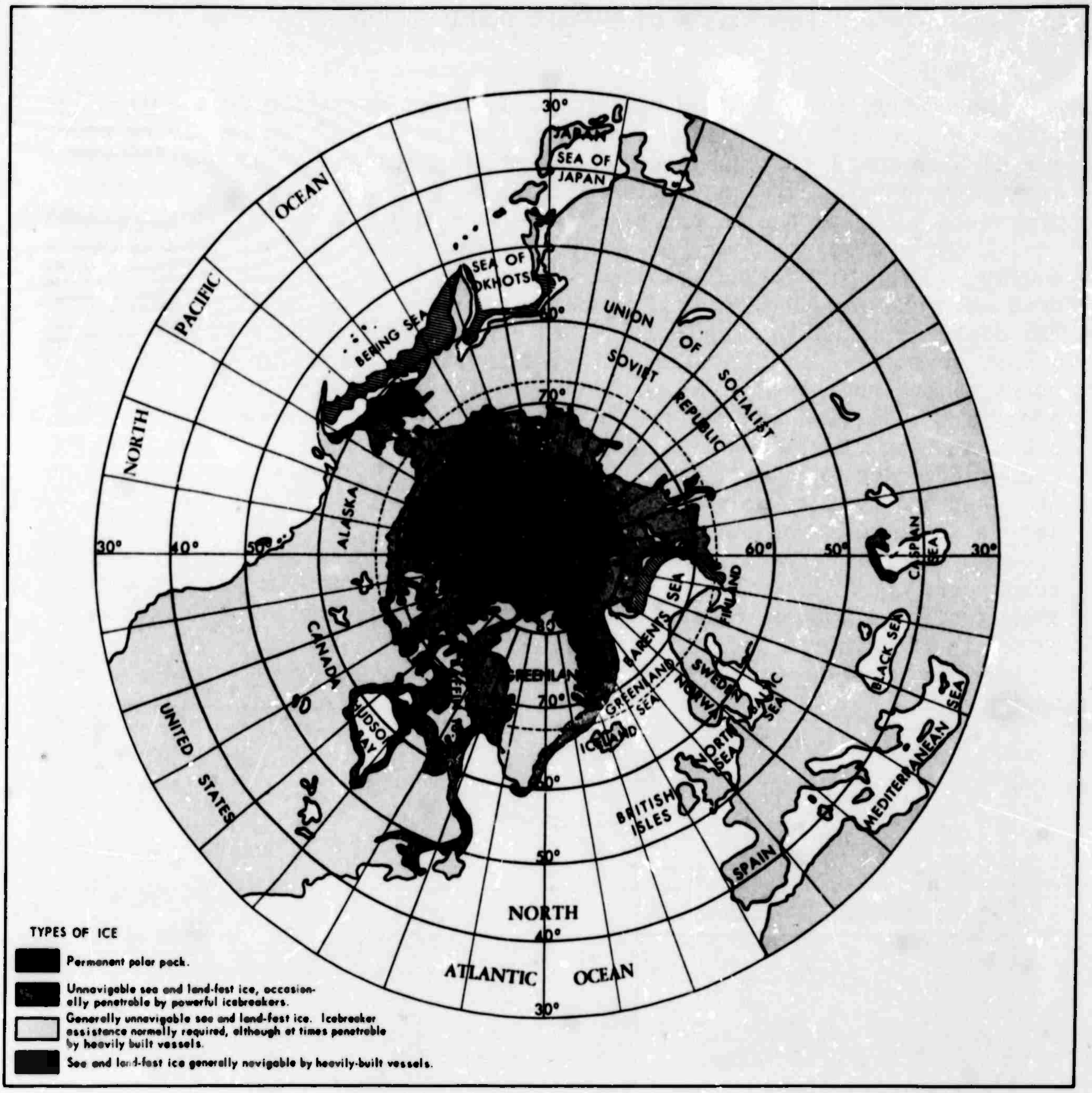


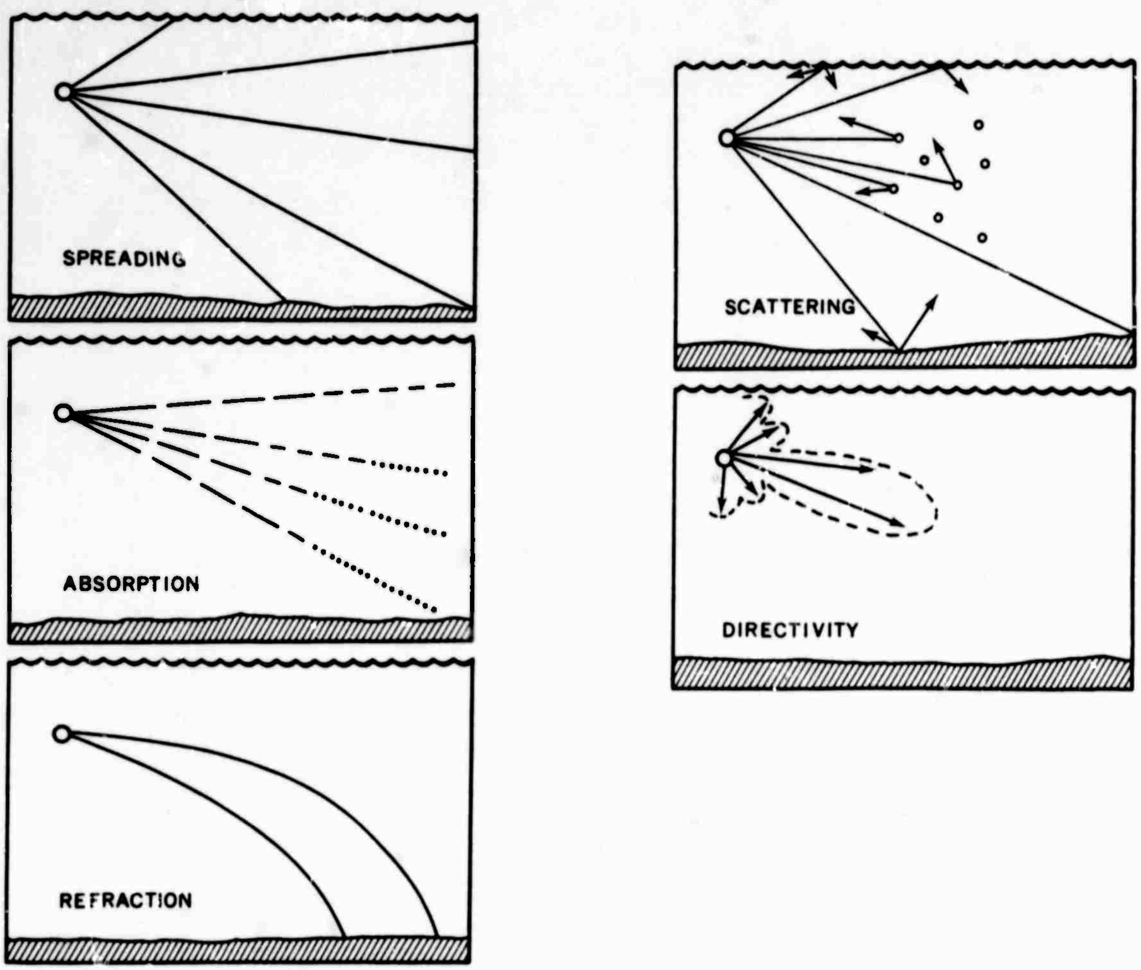
FIGURE 30. AVERAGE LIMITS OF VARIOUS DEGREES OF NAVIGABILITY OF ICE IN THE NORTHERN HEMISPHERE IN JANUARY

PHYSICS OF UNDERWATER SOUND IN THE SEA

Any energy source to be used for underwater detection must fulfill 3 basic requirements: 1) Adequate penetrative range, 2) high resolving power, and 3) high speed of transmission. Of all known physical agents, sound appears to be the only one, at present, that fulfills the 3 requirements to probe the ocean to the depths necessary for modern military requirements.

Sound can be utilized for underwater detection if there is a sound source, a transmitting medium, and a sound detector. The acoustic systems used are echo ranging (active) sonar and direct listening (passive) sonar. The distinguishing features of the two systems are, briefly: The active system employs an echo ranging transducer which emits a sound pulse and receives the reflected sound or echo from a target; the passive system employs a listening device (hydrophone) and requires that the target be the sound source. The effectiveness of these sonar systems is influenced by such factors as equipment design, operator ability, the noise inherent in the ship and system, ship speed, target size and aspect, and the characteristics of the environment.

As a result of environmental barriers in the oceans, the performance of sonar equipment is variable, and detection ranges often vary considerably from predicted values. Some of the factors that must be considered are shown in the diagram below.



Present-day sonar platforms may be ocean bottoms, torpedoes, towed bodies, submarines, surface ships, or aircraft. The uses of underwater sound, in addition to detection, include communication, navigation, weapon delivery, and submarine countermeasures.

BASIC PRINCIPLES

Underwater sounds are produced in one of 3 basic ways: 1) By percussion, as the striking of a bell, gong, or the bottom of the vessel; 2) by oscillation, as the vibration of a diaphram; and 3) by explosion, as by small bomb or depth charge. Certain man-made noises ordinarily produced in water such as those due to operation of the main engines of a vessel, can be detected by an appropriate listening device. In addition, many detectable noises are made by animals living in the ocean.

The direction of travel of sound waves can be measured either by means of binaural hearing (hearing with two "ears") or by equipment which has directional characteristics similar to those of a directional antenna used in radio. Either method can be used for determining the direction from which general noise is coming, but only the latter is used in sonar equipment for determining direction and distance by reception of an echo from a directional signal, in a manner similar to radar.

Distance can be determined by: 1) Measuring the elapsed time between transmission of a signal and return of its echo, 2) measuring the elapsed time between transmission of a signal and its reception at a second station, 3) measuring the time difference between reception of a signal transmitted through water and one transmitted through air, 4) measuring the difference in phase between two signals or change of phase of a signal when it returns as an echo, or 5) measuring the angle at which an echo is received from a signal produced at another source. The first method, used in sonar and echo sounding equipment is similar in principle to radar. The second method is used primarily in radio acoustic ranging in which underwater sound signals trigger a sonobuoy, which transmits a radio signal to indicate the time of reception of the sound signal. The third method is used in special precision navigation systems. The fourth and fifth methods were used in early forms of echo sounders.

The reader must realize that underwater sound is governed by physical laws; therefore, discussion of these is unavoidable. Let us first consider some of the basic aspects of sound.

Sound may be defined as a periodic variation in pressure, particle displacement, or particle velocity in an elastic medium and is, therefore, a form of mechanical energy. All sound, whether produced by a cowbell or a complicated electronic device, behaves in much the same manner. Sound which originates as a wave motion produced by a vibrating source requires for its transmission an elastic medium such as air or water. For example, consider a piston suspended in one of these mediums. As the piston is energized and moves forward and backward, the medium is compressed on the forward stroke and decompressed or rarefied in the return stroke. Thus, a wave motion or series of compressions and rarefactions moves from the source out through the medium. The wave motions are parallel to their direction of propagation

(transmission) and are designated longitudinal waves. A simple longitudinal wave comprises a single compression and a single rarefaction (Fig. 31). A series of compressions and rarefactions, such as is produced by a sound source, constitutes a longitudinal compressional wave train. The distance between any two corresponding points on the wave curve is the wave length (λ). Since each wave is produced by one complete vibration or cycle, the frequency of a sound wave is defined as the number of complete cycles passing a given point in a specified amount of time and is usually expressed as cycles per second (cps).

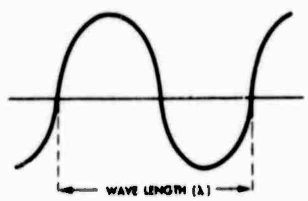


FIGURE 31. LONGITUDINAL SOUND WAVE

As sound energy travels outward from the source into the medium the constant mean pressure at a point assumes a new value known as the instantaneous pressure. This pressure change is detectable by a hydrophone and can be electronically displayed and analyzed. If a single pulse is considered, the course of the sound energy in the medium can be followed by placing recording microphones (or hydrophones) in the general vicinity and noting the time of first response. By using a sufficiently large number of microphones, it is possible to record all those points in space which are reached by the spreading sound pulse simultaneously. The surface on which all of these points are located is called a sound front or a wave front as shown in Figure 32. Energy reaching any point along a wave front will do so along a line called a ray path or a sound ray. A sound ray may be defined as a line everywhere perpendicular to the wave front, and it indicates direction of travel of the wave.

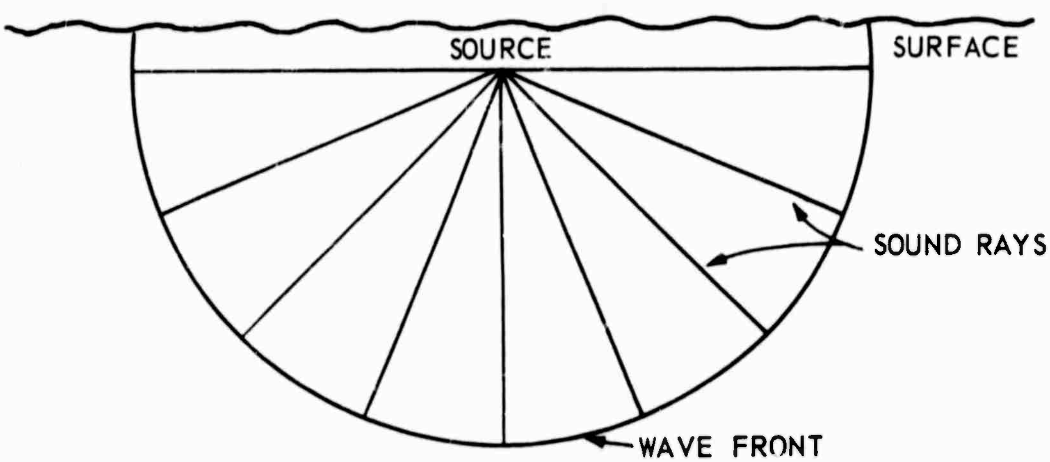


FIGURE 32. WAVE FRONT

INTENSITY AND PRESSURE LEVEL

Sound intensity can be defined in a number of ways, but for practical purposes it usually is defined as: $I \propto p^2$ ($dynes^2/cm^4$). (Intensity is proportional to the square of the pressure.) I is the intensity of sound, and p is sound pressure at the source. Dynes per square centimeter are the units ordinarily encountered in oceanography. One dyne per square centimeter is equal to one microbar. The proportional relationship is convenient because pressure is easily measured with pressure sensitive hydrophones whose voltage outputs are proportional to pressure.

The values of pressure encountered in practice can vary from 10^{-4} to over 10^6 dynes/cm^2. Because of this large variation, it is convenient to use a logarithmic scale to the base 10 to specify sound pressures. The fundamental unit is the bel. The decibel (db), a smaller and more convenient unit, is one-tenth of a bel. The pressure level (L) in decibels is a ratio and is defined as:

$$L = 10 \log \left(\frac{p}{p_o}\right)^2 = 20 \log \left(\frac{p}{p_o}\right),$$

where p is the sound pressure measured at the hydrophone, and p_o is a reference sound pressure. The reference sound pressure serves to orient the starting point for the scale of pressure levels and usually is either 1 dyne/cm^2 or 0.0002 dyne/cm^2 *. Assuming p_o to be unity, the equation simplifies to $L = 20 \log p$. Doubling a sound pressure corresponds to an increase of sound pressure level of 6 db. Sample computations of pressure level are:

(a) Pressure (p) = 1,000 dynes/cm^2, reference level (p_o) = 1 dyne/cm^2, $L = 20 \log (1000/1) = 60$ db (relative to 1 dyne/cm^2).

(b) Pressure (p) = 1,000 dynes/cm^2, reference level (p_o) = 0.0002 dyne/cm^2, $L = 20 \log (1000/0.0002) = 20 \log 1000 + 20 \log 5000 = 60 + 74 = 134$ db (relative to 0.0002 dyne/cm^2).

Because it is sometimes difficult to associate sound pressure levels of common noises with numerical values the following table is presented (values in decibels):

* In this publication, 1 dyne/cm^2 will be used exclusively as the sound reference pressure, but the reader must be cautioned always to check the reference pressure in other sources.

Sound Pressure Level Relative to 0.0002 dyne/cm^2	Sound Pressure Level Relative to 1 dyne/cm^2	Description
120	46	Pneumatic drill at operator's position
100	26	Loud motor horn at 23 feet
80	6	Very heavy street traffic (New York)
70	-4	Loud peal of thunder
50	-24	Conversational voice at 12 feet
30	-44	Quiet suburban street
10	-64	Rustle of leaves
0	-74	Faintest audible sound

From the above table it may be noted that to change from a reference level of 1 dyne/cm^2 to 0.0002 dyne/cm^2 all one need to do is add 74 db.

The relationship between pressure and intensity of sound in decibels relative to 1 dyne/cm^2 is

$$L\ (\text{db}) = 20 \log P = 10 \log I$$

Each decibel increase in level represents a 13% increase in sound pressure or a 26% increase in sound intensity. The overall intensity of sounds from any source, and the manner in which the sound energy is distributed over a frequency spectrum may fluctuate widely with time. For analysis, spectrum level graphs for any sound can be constructed which show frequency (usually in a band 1 cps wide) versus db level, or an overall level of sound may be computed which includes all frequencies. Analysis by frequency octaves (or fractional octaves) provides a fast method of processing and displaying acoustic data. When a frequency band under consideration has a band ratio of 2:1, then the bandwidth is called an octave ($f_2 = 2f_1$). For example, if f_1 equals 1000 cps, then f_2 equals 2000 cps, and the octave includes all frequencies between 1000 and 2000 cps. Figures 33 through 37 show the effect of using bandwidths to analyze sound. The curves are all plots of the same noise, but differ radically when subjected to bandwidth changes.

SPEED OF UNDERWATER SOUND

Three variables govern the speed (c) of sound in a fluid. They are density (ρ), compressibility (β), and the ratio between the specific heats of the fluid at constant pressure and at constant volume (γ). The following formula is applicable:

$$c = \sqrt{\frac{\gamma}{\rho\beta}}\ .$$

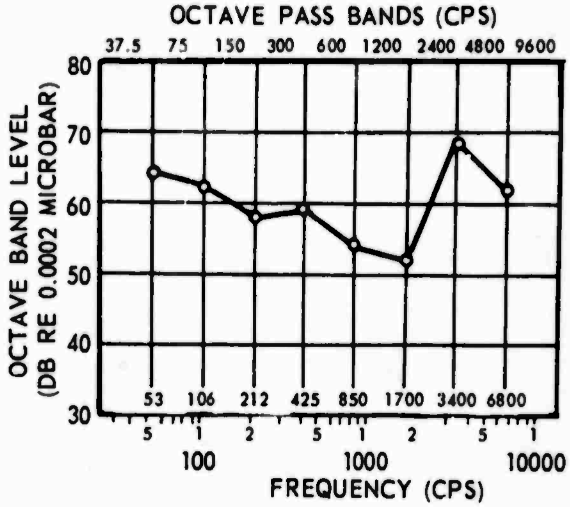


FIGURE 33. EXAMPLE OF PLOT OF NOISE SPECTRUM BY OCTAVE BANDS

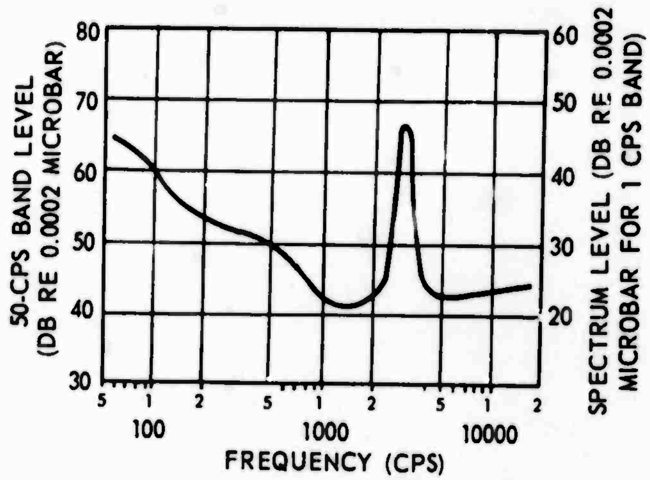


FIGURE 36. THE SAME NOISE PLOTTED WITH A CONSTANT BANDWIDTH (SOCPS) ANALYZER

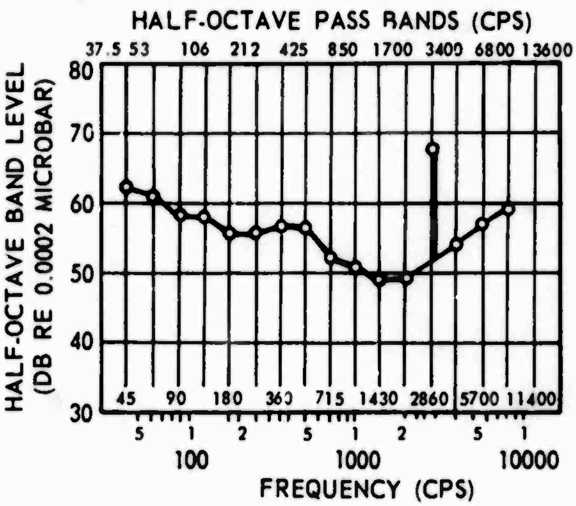


FIGURE 34. THE SAME NOISE PLOTTED BY HALF-OCTAVE BANDS

(R. W. YOUNG, 1955)

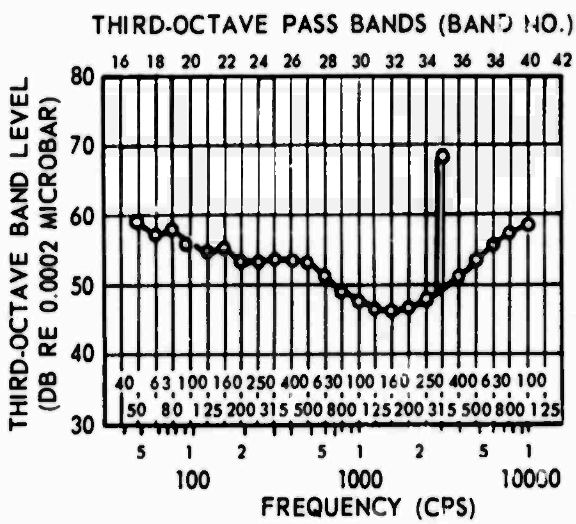


FIGURE 35. THE SAME NOISE PLOTTED BY THIRD-OCTAVE BANDS

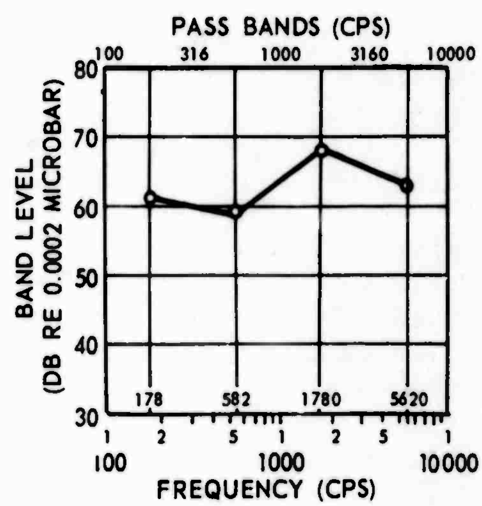


FIGURE 37. THE SAME NOISE PLOTTED BY BANDS AVAILABLE ON A SOUND ANALYZER

Density is mass per unit volume.

$$\rho = \frac{\text{Mass}}{\text{Volume}} \quad \text{and Mass} = \frac{\text{Weight}}{\text{Gravity}} \quad \text{so}$$

$$\rho = \frac{\text{Weight}}{\text{Volume. Gravity.}}$$

Compressibility refers to the relative change in volume for a given change in pressure and is the inverse of bulk modulus. The compressibility of water is low, and consequently the speed of sound in water is high. The specific heat ratio enters the formula because the energy of a sound impulse is briefly transformed into heat, and then reconverted (with slight loss) into kinetic energy. The ratio rarely exceeds 1.02 in sea water and is commonly taken as unity. Note that the speed of sound in sea water does not vary with changing frequency. The same formula can be used to determine the speed of sound in air. The speed of sound in water is approximately 4.5 times the speed in air.

An increase in temperature decreases both density and compressibility, resulting in an increase in the speed of sound. In sea water, an increase in pressure or salinity produces a slight increase in density and a larger decrease in compressibility, resulting in a net increase in the speed of sound. Thus, in sea water, an increase in temperature, salinity, or pressure increases sound speed. Of the three, temperature has the greatest influence on the speed of sound in sea water. The pressure effect is slight, and the change of salinity is not sufficiently great to exercise a marked influence on the speed.

The composite effect on sound speed can be shown as follows (from Kuwahara):

$$c = 4422 + 11.25\,T - 0.045(T^2) + 0.0182\,D + 4.3(S-34)$$

where c = sound speed (feet/second),

T = temperature (°F),

D = depth (feet), and

S = salinity (‰).

More exact equations (Wilson's equations in NAVOCEANO H.O. SP-58) are in use to determine sound speed; however, the above equation gives a good approximation.

Normally the change of these 3 physical properties is much more rapid in a vertical direction than in a horizontal direction. The change with depth varies with location. With respect to temperature, much of the ocean is considered to consist of 3 layers, a surface layer influenced greatly by the temperature of the air above it, a thermocline of rapidly decreasing temperature, and a nearly uniform deep-water layer. The increase of pressure

with depth is almost uniform, the pressure at 3,000 meters being approximately twice that at 1,500 meters, and ten times that at 300 meters. Typical profiles of temperature, salinity, and speed of sound with depth are shown in Figure 38.

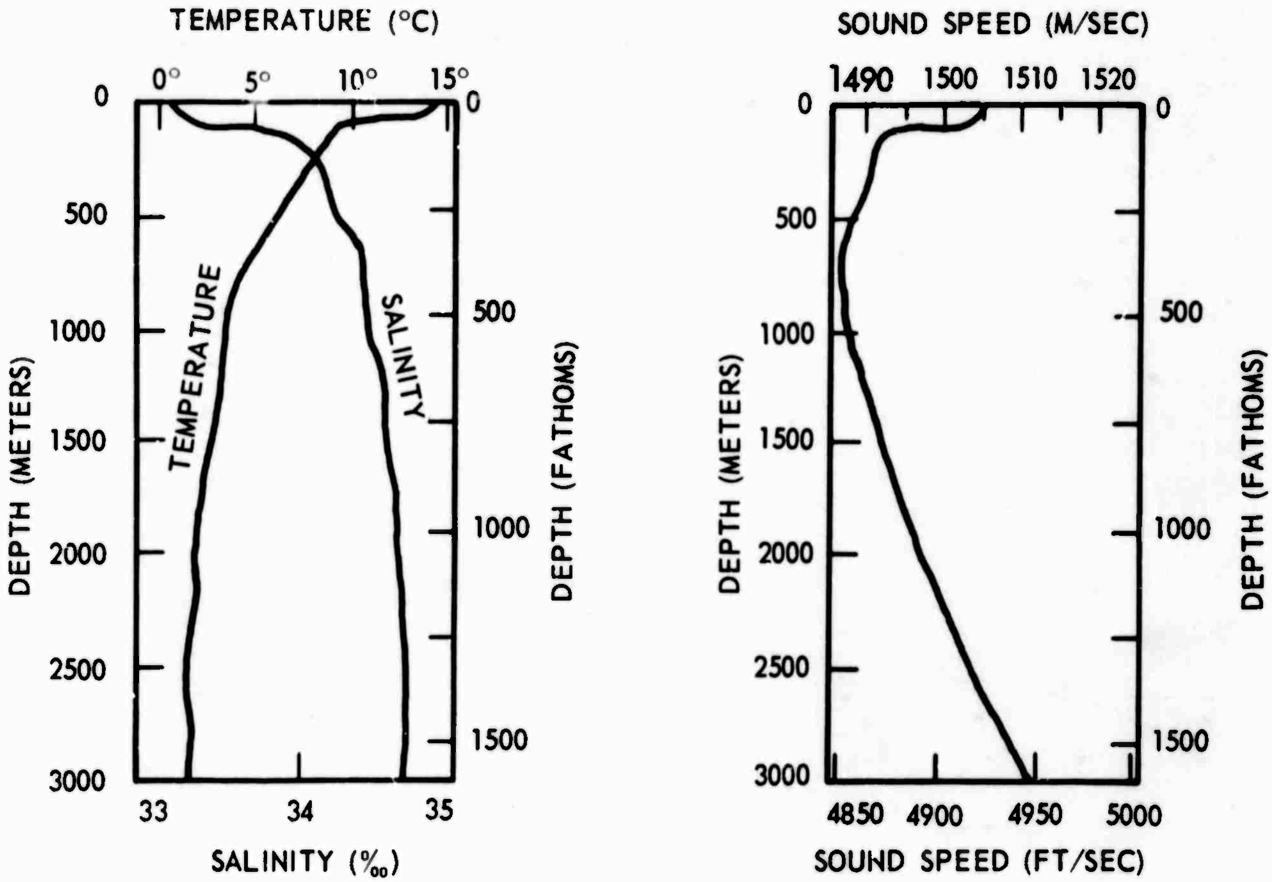


FIGURE 38. TYPICAL PROFILES OF TEMPERATURE, SALINITY, AND SOUND SPEED

Once the sound speed in the medium has been determined, the wavelength of sound at different frequencies can be found by dividing the speed by the frequency. At low frequencies the wavelength is very long, whereas at high frequencies the wavelength is very short. For example, at 4 cps the wavelength is about 366 meters, and at 40,000 cps the wavelength is about 3.6 cm (at a sound speed of 1463 m/sec).

Study of transmission of sound from underwater explosions indicates that near the explosion the speed of sound may be somewhat higher than expected, probably due to increased pressure caused by the disturbance. This effect extends over such a short distance that it is insignificant in ordinary underwater sound transmission studies.

TRANSMISSION LOSS

If the ocean were infinite in extent (boundless), nonabsorbing, and its physical properties unvarying, sound would travel at a constant speed and radiate outward in straight lines or rays. If we examine only a small cone of about 30°, we observe that as the sound spreads, the cross sectional area of the cone increases. The total energy or power radiated is constant in each of the conic sections, but the energy is spread over an ever increasing area as the distance (r) from the source increases. The intensity of sound (I) at any point in the cone decreases as the inverse square of the distance $\left(I = \frac{I_1}{r^2}\right)$ where I_1 is intensity one yard from the source. This is known as the inverse square law. The losses resulting from the inverse square law are called spherical spreading losses and are about 6 db per distance doubled. For example (see Fig. 41), at 100 yards range the loss is 40 db. Doubling the range to 200 yards results in a 6 db increase in loss or a total loss of 46 db. In Figure 39, the energy passing through the small area at cone A spreads outward, becoming less intense per unit area at cone B, and so on.

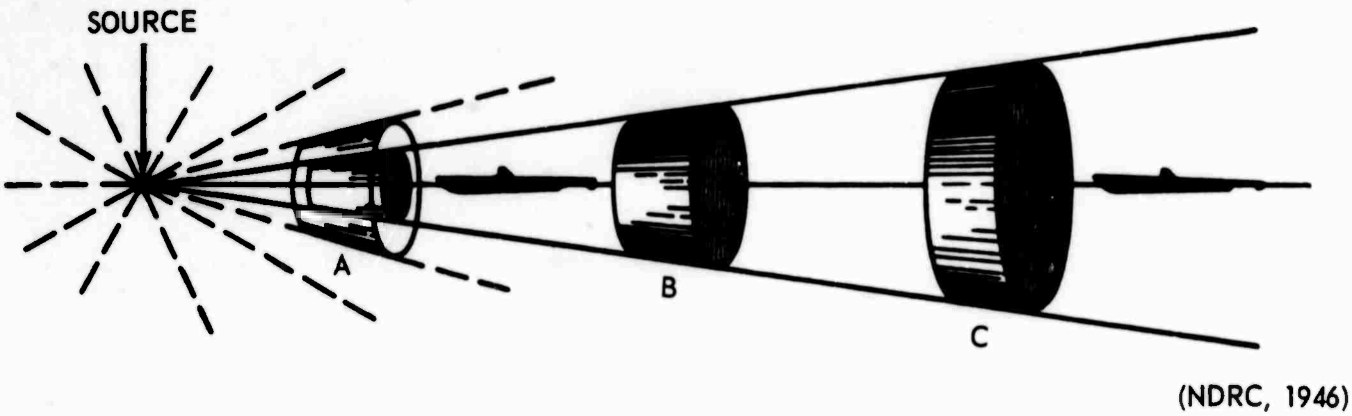


FIGURE 39. DIVERGENCE OF SOUND RAYS FROM OUTGOING PING

Losses are computed relative to a point one yard from the sound source. Suppose for example, in Figure 40, A is one yard, B is 1,000 yards, C is 5,000 yards, and the intensity is desired at points B and C relative to A. The inverse square law would determine the intensity of sound at B as 1/1,000,000 of that at A, while at C it is 1/25,000,000 of that at A. Numbers such as these are awkward to use; therefore, in practice two sounds of different intensities usually are compared on a logarithmic scale. This scale tends to narrow the numerical range between very faint and very loud sounds and replaces multiplication and division by addition and subtraction in sonar equations. To find the decibel difference between points A and B in the example cited, the common logarithmic (base 10) of the ratio A:B is multiplied by 10 $\left(L = 10 \log \frac{I_A}{I_B}\right)$. Thus, if the intensity at A is 1,000,000

times the intensity at B, 10 times the log 1,000,000 is 60 and the intensity level at A is 60 db above that at B. Transmission losses may be calculated by considering either pressure levels or intensity levels by using the following equation:

$$L = 10 \log I = 20 \log P \text{ (answer in db relative to one dyne/cm}^2\text{)}.$$

Pressure levels are commonly used in underwater sound because hydrophones respond directly to pressure variations of the sound wave.

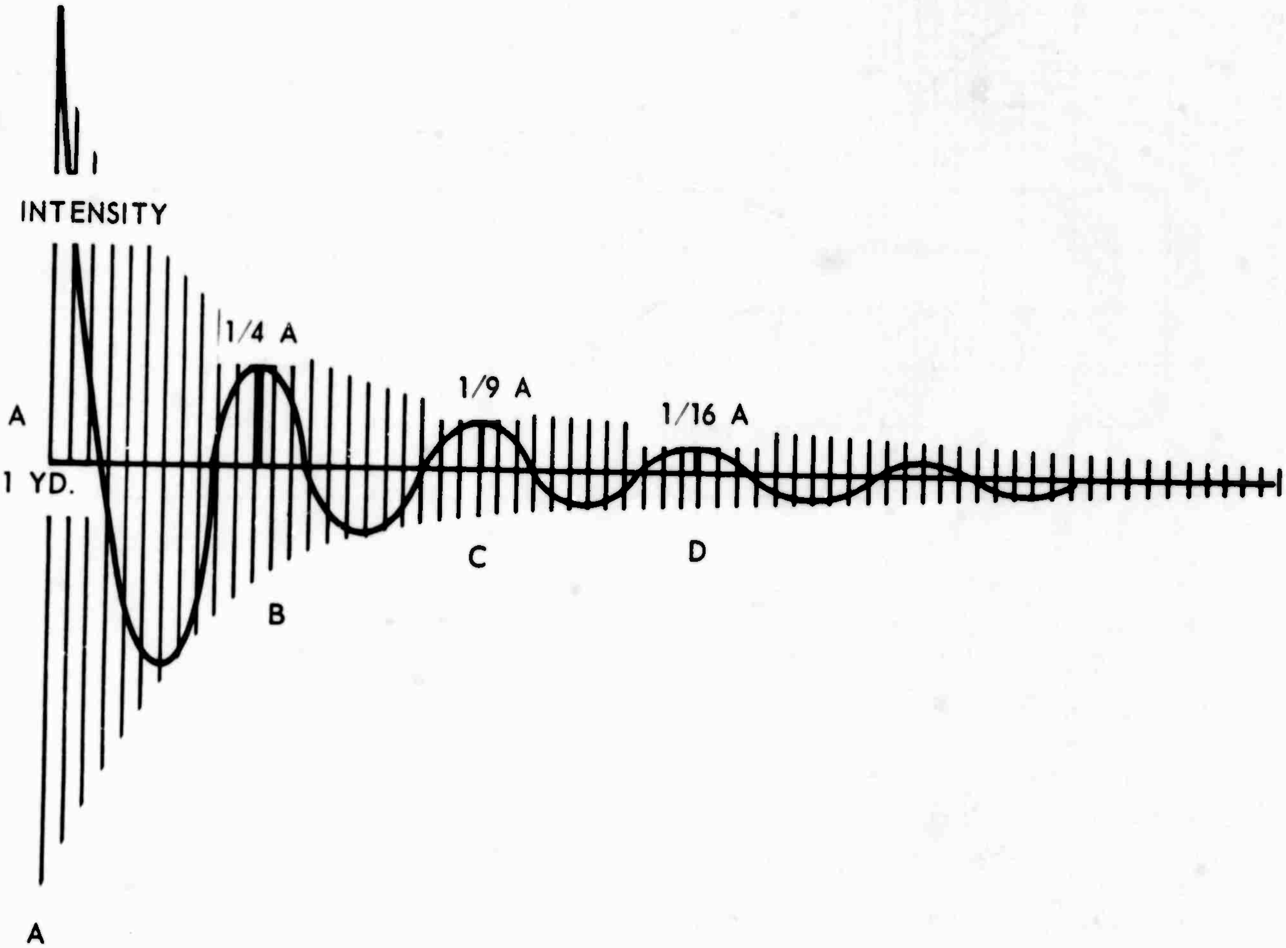


FIGURE 40. RELATIVE DECAY OF INTENSITY DUE TO SPHERICAL SPREADING

Because the ocean is not boundless, the inverse square law alone will nct always apply; at extreme distances from the source and in sound channels where the energy is prevented from spreading further vertically, sound pressure level decreases as the inverse first power of distance, or about 3 db per distance doubled. This is called cylindircal spreading loss. A third type of spreading occurs when the sound pressure level decreases as the

inverse fourth power of distance, or about 12 db per distance doubled; this is known as dipolar spreading loss and occurs generally in shallow water. Some combination of these three types can be experienced whenever sound is transmitted in the ocean. A comparison of loss curves as a function of distance for the 3 types is shown in Figure 41.

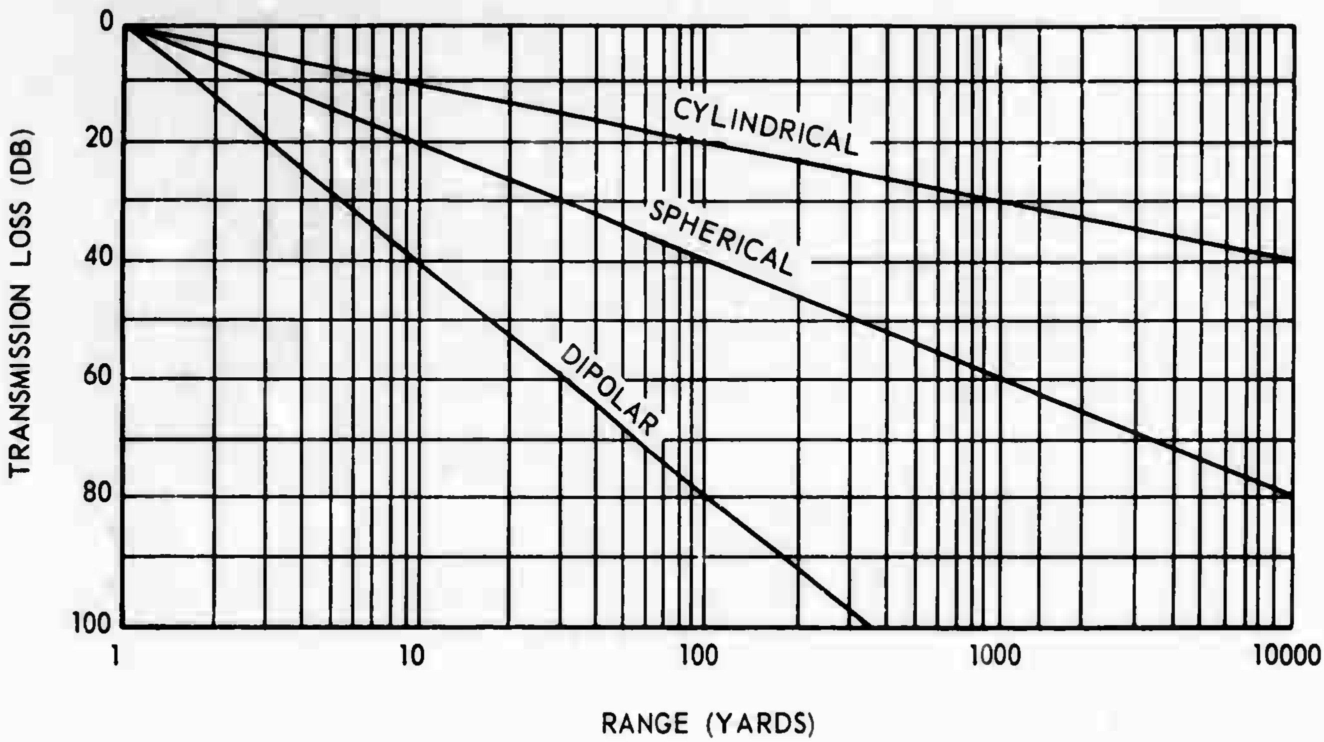


FIGURE 41. COMPARISON OF SPREADING LOSSES

When sound travels through sea water, additional losses occur from: 1) Attenuation (deflection and scattering by suspended particles and absorption by generation of heat as sound passes through the water), 2) reflection at the boundaries, and 3) refraction because of sound speed gradients. Transmission of sound is a function of the oceanographic environment and the frequency of the sound source. The general effects on transmission loss ascribed to spreading and attenuation as a function of frequency and range are shown in Figure 42. The straight line at the top of the figure indicates the loss from spherical spreading alone and shows that the sound pressure level falls off at 6 db per distance doubled. The other curves show the transmission loss at different frequencies resulting from both spreading and attenuation. For ranges less than or equal to 1,000 yards, little variation in loss occurs as the frequency changes. Note, however, that smaller losses and longer ranges are achieved at the lower frequencies.

ATTENUATION

In addition to spreading loss, sound energy propagated through a volume of sea water undergoes some loss of energy because of attenuation, that is,

absorption and scattering. In the passage of sound through water, some of the energy is converted into heat; this is called absorption. Scattering loss results from reflectors in the water that may vary in size from minute particles to large objects.

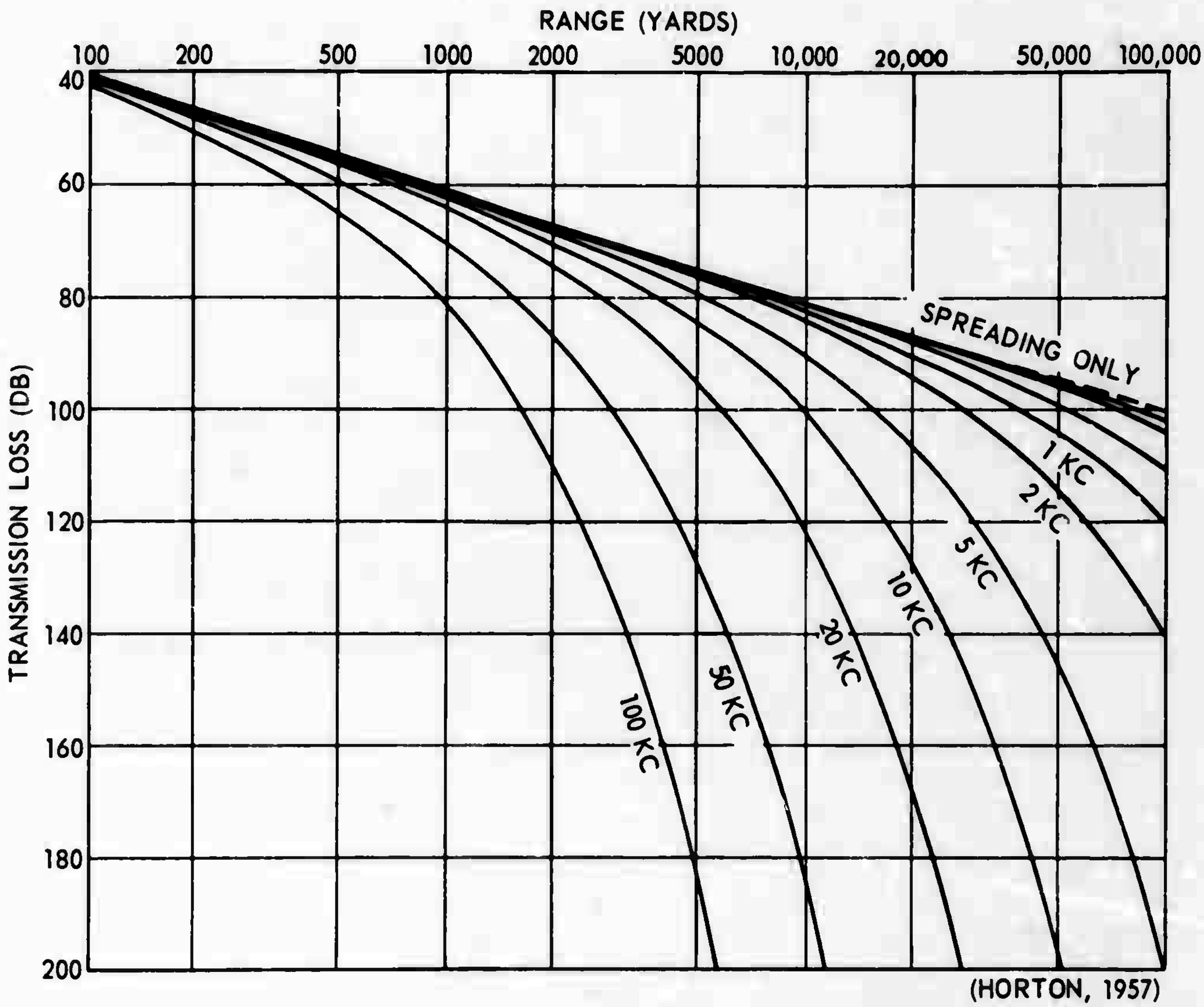


FIGURE 42. SPHERICAL SPREADING AND ATTENUATION LOSS AS RELATED TO FREQUENCY AND RANGE

Absorption is caused by a number of factors:

(1) The viscosity of the medium converts sound energy into heat by internal friction. Viscosity causes high losses.

(2) The sound waves alternately raise and lower the temperature of the water by a small amount. Because of this thermal conduction, some sound energy is converted to heat and is lost.

(3) Sound waves set suspended particles or organisms oscillating which dissipates some of the sound energy in the form of heat. This process is especially significant if the suspension contains air bubbles.
(4) Heat loss is produced by dissociation of dissolved salts in the water caused by sound energy.
(5) At the water-bottom interface some sound energy is absorbed by certain sediments. At the air-surface interface a small percentage of absorption occurs depending upon sea state and other factors.

While heat loss is insignificant as far as the warming of the ocean is concerned, it is not negligible when viewed in terms of the energy remaining in a sound wave after it has passed through an appreciable amount of water.

Absorption loss usually is expressed in terms of the absorption coefficient, which is a value representing the amount of acoustic energy absorbed per unit distance. Absorption at high frequencies is much greater than at low frequencies. Figure 43 shows curves of absorption coefficient as a function of frequency as formulated by various workers in the field.

Although volume scattering is a component in the attenuation of sound, its contribution is not as important as that of absorption. As a sound wave passes an obstacle suspended in the medium, the obstacle is set into vibration and becomes a secondary source of sound. The amplitude and frequency of the vibration is proportional to the amplitude and frequency of the primary sound. The obstacles may be microscopic air bubbles or objects as large as whales. Most of the sound travels on unaffected by these obstacles, but some of it is reflected (scattered) in various directions. Much of this scattered energy may go off at angles of only 1 or 2 degrees from the original direction, but a small part goes off in widely different directions. Sound energy will also be scattered at the surface or bottom provided the surface is disturbed by waves and the bottom is irregular. The part of the scattered ray returned to the receiver is known as reverberation; the part not returned is lost energy.

The behavior of small solid particles as scatterers of sound in sea water is dependent on the area of the particle that effecitvely reflects sound. The larger this area, the more efficient the particle is as a scatterer. This efficiency may be measured roughly by the ratio of the circumference of the particle to the wavelength of the sound beam raised to the fourth power, that is, $(\frac{\pi D}{\lambda})^4$. The scattering power of the particle is increased when: 1) The circumference is increased, 2) the wavelength is reduced, or 3) both occurs simultaneously. When the ratio of the circumference to the wavelength is greater than one, the reflecting area approaches that of the scatterer, and its scattering efficiency is increased. However, when the ratio is less than one the reflecting area is greatly reduced, and thus its scattering efficiency is also reduced. This is demonstrated in Figure 44. Moreover, the same object scatters more of higher frequency sounds than lower frequencies. The wavelength of a 2-kc sound source is ten times that of a 20-kc source, According to the ratio, therefore, a sound scatterer will scatter 10,000 times more sound for a 20-kc source than for a 2-kc source.

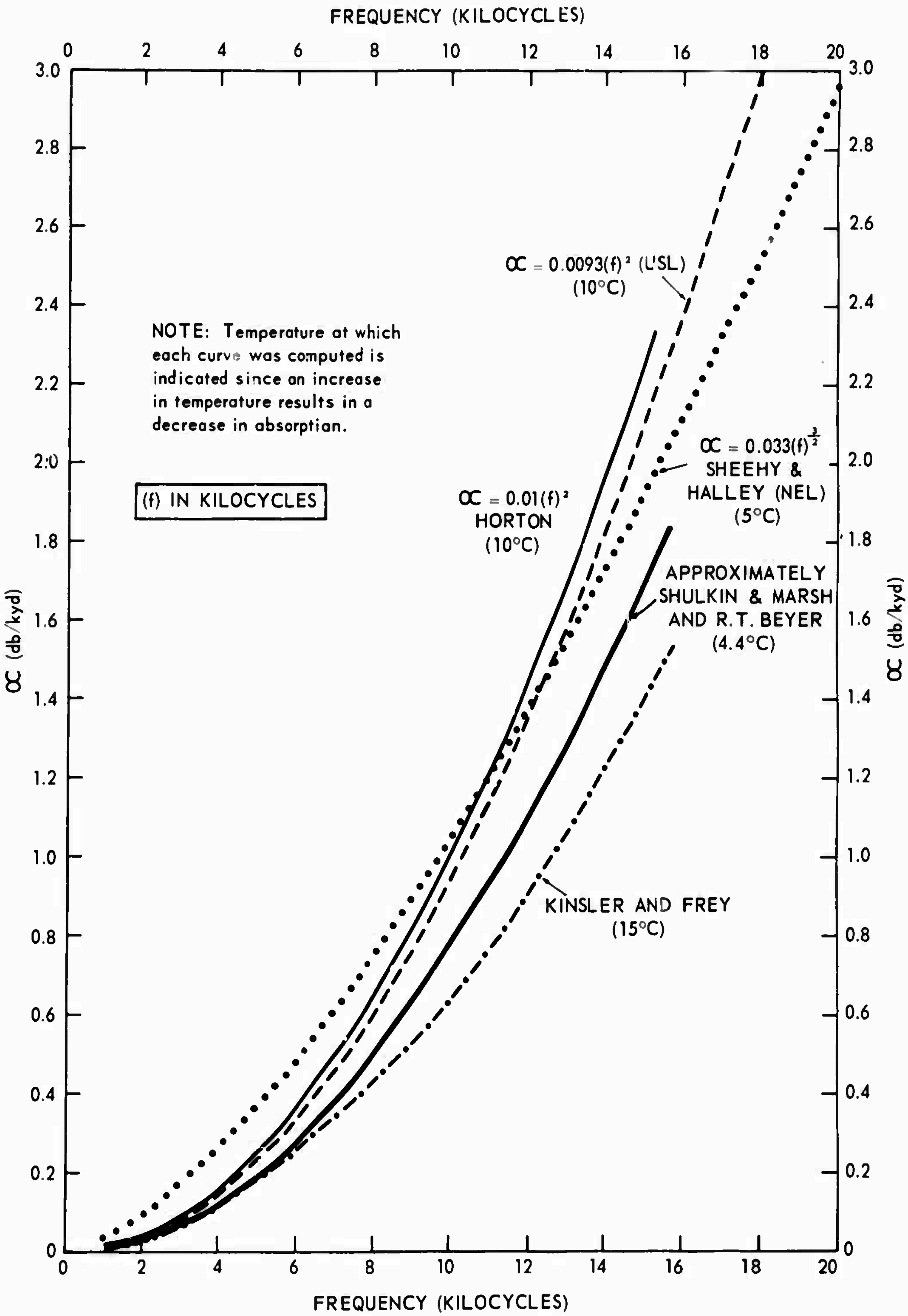


FIGURE 43. ABSORPTION COEFFICIENT IN THE WATER COLUMN AS A FUNCTION OF FREQUENCY

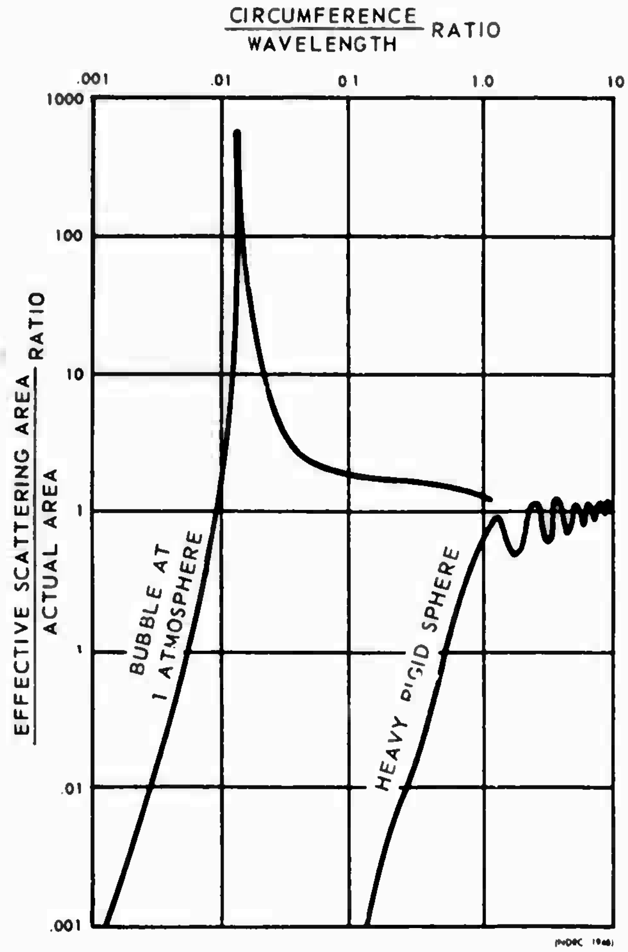


FIGURE 44. VARIATION OF EFFECTIVE SCATTERING AREA WITH WAVELENGTH

Although the scattering ability of bubbles (produced by such action as surface waves, ships, torpedoes, submarines, and marine organisms) also is dependent on reflective area, their acoustic behavior differs from that of solid particles. The scattering power of bubbles increases with a decrease of circumference/wavelength ratio values to about 0.012 and then falls off rapidly for lesser ratio values (Fig. 44). This characteristic of bubble scattering is related to the resonant frequency of the bubble which depends on the diameter and the inside pressure of the bubble. When sound approaching the resonant frequency of a bubble impinges on it, the sound produces a high amplitude vibration of the bubble, and, consequently, the reflective cross section becomes very large. For example, a 1-inch bubble at the surface is found to have a resonant frequency of 0.12 kc, and this is equivalent to the circumference/wavelength ratio of 0.012. Thus at the ratio value of 0.012, a 1-inch bubble at atmospheric pressure has its maximum reflective area. Figure 44 shows that gas bubbles scatter low-frequency sound more effectively than a small solid scatterer of the same size.

The preceding paragraphs assume spherically shaped scatterers. Any deviation from this shape produces corresponding changes in reflective area calculations. Moreover, the reflective properties of sound scatterers are

not too well known, and the determination of scattering ability is made more difficult.

Sea ice effectively scatters sound beams, especially when the underwater profile of the ice is uneven, as in hummocky ice fields and with ice containing numerous air bubbles. In ice fields the low frequencies (less than 30 cps) appear to be transmitted most effectively owing to the scattering of the higher frequencies (particularly those around 22 kc).

REVERBERATION

When a sound pulse is emitted in the ocean, the echo from a target may be masked by interfering or unwanted sounds. This background of interference consists of both reverberation and background noise. Reverberation is distinguished from background noise in that reverberation results directly from the emitted sound pulse.

If the surface and bottom of the ocean were absolutely smooth and no suspended matter (including fish) were in the water, there would be no reverberation. However, irregularities in the ocean surface, bottom, and the water itself are capable of scattering the sound pulse and echoes. That portion of the scattered sound which returns to the transducer is called reverberation. Reverberation may therefore be considered the resultant of a large number of very weak unwanted echoes. The intensity with which reverberation occurs is directly proportional to the source intensity, ping length, and beam width. Thus, increased sound output results in increased reverberations; a long ping length causes more reverberation than a short one; and a wide beam increases reverberation. Since wanted echoes become fainter with increasing time interval between ping emission and echo reception, echoes from a distant target that are audible over other background noise may be masked by reverberation. Thus, the echo-to-reverberation ratio is more important than the reverberation itself. The echo and the reverberation are interrelated in that each is a sound of definite pitch. The difference in pitch (frequency change) between the two, called doppler effect, results from the motion of the target relative to its surroundings. Doppler makes the echo recognizable over a background of reverberation by giving the echo a different pitch from that of the reverberation. The larger the doppler effect the fainter the echo that can be recognized. For instance, if a target is moving at 15 knots directly toward the searching ship, an echo obtained at high frequencies can be heard even though it may be 15 to 20 db below the reverberation.

The ocean contains many kinds of suspended particles capable of returning an echo. These echoes may be too weak to be heard individually, but collectively they can reinforce each other and return to the sound projector as reverberation. Scattering from the surface, from the bottom, and from particles in the water is called surface, bottom, and volume reverberation, respectively.

Surface Reverberation - For short ranges, surface reverberation increases with wind velocities between 7 and 18 knots as waves and mixing increase. Above 18 knots, an acoustic screen is effectively formed near the surface by entrapped bubbles, preventing any further increase in reverberation level.

Surface reverberation from ranges in excess of 1,500 yards usually is lower in level than either bottom or volume reverberation.

Bottom Reverberation - This type of reverberation is of primary importance in shallow water. It is greatest over rock, shell, cobble, or rippled sand bottoms; it is least over mud bottoms. The most intense reverberations comes from rock and coral and is about 10 times as loud as reverberation from mud. Reverberation is also a function of the grazing angle (angle that the sound ray makes with the bottom). When negative sound speed gradients exist, downward refraction causes larger grazing angles in shallow water and raises the reverberation level.

Volume Reverberation - The intensity of this type of reverberation of sound depends upon the number and distribution of scatterers, as well as their size, shape, and reflectivity. If the density of these reflectors were constant, volume reverberation would decrease as the inverse square of the range. In other words, it should decrease 20 db for each tenfold increase in range. In practice, however, no such uniformity exists. Departures from this rule are ascribed, among other things, to the presence of the deep scattering layer (DSL). The presence of this layer causes an average increase in the reverberation level of about 10 db.

REFLECTION OF UNDERWATER SOUND WAVES

In water, as in air, sound is reflected by obstructions in the form of solid objects or sharp discontinuities. Thus, sound is reflected from the bottom, the shore, hulls of ships, the surface of the water, etc. It is this reflecting energy that is used in echo sounders to determine depth, and in sonar equipment for echo ranging. Sound energy striking some solid surface: 1) May be reflected as a mirror reflects light with little loss of intensity, 2) may be scattered in many directions, 3) may be lost by absorption into the medium, or 4) some combination of the above.

The surface of the sea is rarely smooth; therefore, sound energy striking it is seldom reflected specularly (mirror reflection). Instead, only minute elements of the sea surface reflect sound as a mirror; however, because the orientation of each of these "mirrors" is changing continuously, the sound energy is reflected in many directions (reverberation). Aside from the need for an accurate description of the sea surface, prediction of the reflective characteristics of the sea surface necessarily includes frequency of sound source and grazing angle. Figure 45 illustrates the increase in surface scattering with an increase in wind speed and grazing angle.

Because the acoustic impedance ρc (density times sound speed) mismatch between air and water is so great, almost all sound energy is reflected at the air-sea interface (Fig 46). The fact that 99% of the sound energy in echo ranging is confined to the sea is important because: 1) Longer surface duct ranges are obtained, and 2) some portion of the sound energy is reflected back to the source as reverberation.

The ocean bottom also may reflect sound; however, the amount of sound energy reflected from the bottom depends upon the type of bottom material,

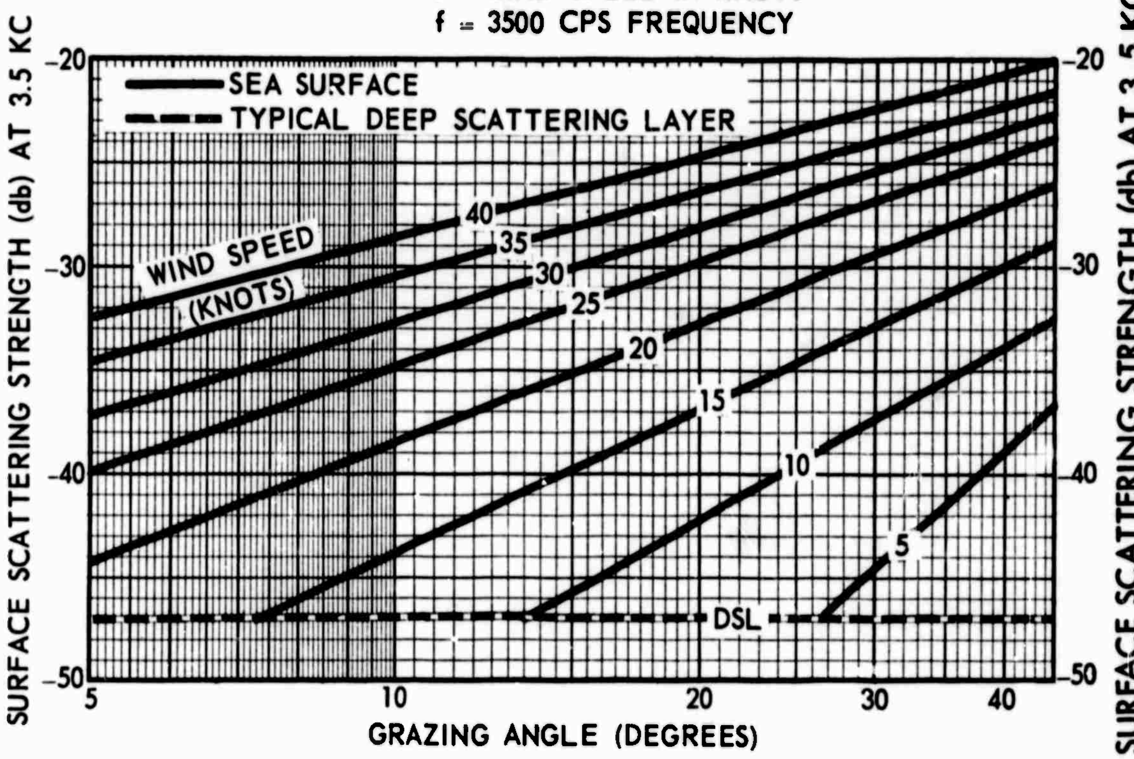
grazing angle, and frequency. A smooth sand bottom reflects sound very effectively. In contrast, a soft mud bottom is an especially poor reflector. A smooth rock bottom is perhaps the best reflector because the amount of energy absorbed (high ρc) is small. Unfortunately, in many areas rock bottoms are irregular and consequently reflect sound in many directions. Much of this energy is scattered back to the sound source causing intense reverberation which effectively masks any possible target signal. In echo sounding, a layer of soft mud over rock may result in two echoes, indicating two depths.

SEA SURFACE SCATTERING STRENGTH PER UNIT AREA (10 log S) =

$$3.3\,\beta \log \frac{\theta}{30} - 42.4 \log \beta + 2.6$$

$$\beta = 158\,(v f^{1/3})^{-0.58}$$

WHERE θ = GRAZING ANGLE
v = WIND SPEED IN KNOTS
f = 3500 CPS FREQUENCY



(CHAPMAN & HARRIS, 1962)

FIGURE 45. SCATTERING STRENGTH OF 3.5 KC SOUND FROM THE SEA SURFACE AND DSL

Figure 47 shows how refraction and bottom reflection contribute to ranges in shallow water. Figure 47A shows that refraction is slight and much of the sound is transmitted directly over long ranges. Figures 47B and C show that with a strong negative temperature gradient, refraction and bottom reflections result in considerably longer ranges.

Obviously bottom reflection is most effective over reasonably flat bottoms. When echo ranging occurs over hummocky or steeply sloping bottoms, ranges are adversely affected. For example, a submarine might hover behind a rise on the bottom in shallow water where little if any sound energy could reach it. Also, when sound speed gradients cause downward refraction over a sloping bottom (Fig. 48), submarines maneuvering at moderate depth and range may successfully avoid detection. As can be seen from Figure 48 the sound beam bounces down the slope resulting in considerably shortened ranges except for targets located on the bottom.

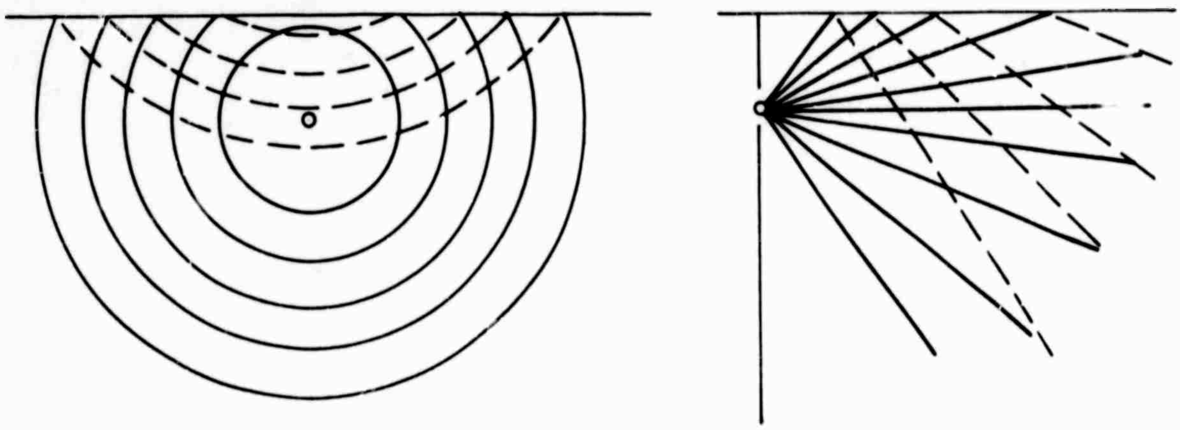


FIGURE 46. SURFACE REFLECTIONS IN ISOTHERMAL WATER

Fish and even tiny sea animals also reflect sound. As a result, echo sounders are widely used among fishermen to locate schools of fish. In deep water it is not unusual for an echo sounder to receive an echo from a depth of about 200 fathoms, although the depth increases somewhat at night. This phantom bottom or deep scattering layer, which is undoubtedly the source of many erroneous shoal sounding reports, is believed to be due to large numbers of tiny marine animals. A sharp discontinuity within the water also causes reflection of sound. Thus, an echo sounder may detect the boundary between a layer of fresh water overlying salt water; a condition which might occur near the mouth of a river.

REFRACTION OF UNDERWATER SOUND WAVES

The laws of refraction as applied to light and radio waves apply also to sound. Because of differences of speed of sound in sea water, an advancing sound wave is refracted toward the area of slower speed. If sound is traveling vertically downward, as in echo sounding, the effect of refraction is relatively slight because the layers of water in which sound speed differ are approximately horizontal, and when the direction of travel of the sound is normal to the refracting surface or layer, there is little refraction.

When a beam of sound is directed in a horizontal direction, however, refraction is greatest. If the sound speed decreases with depth, the upper

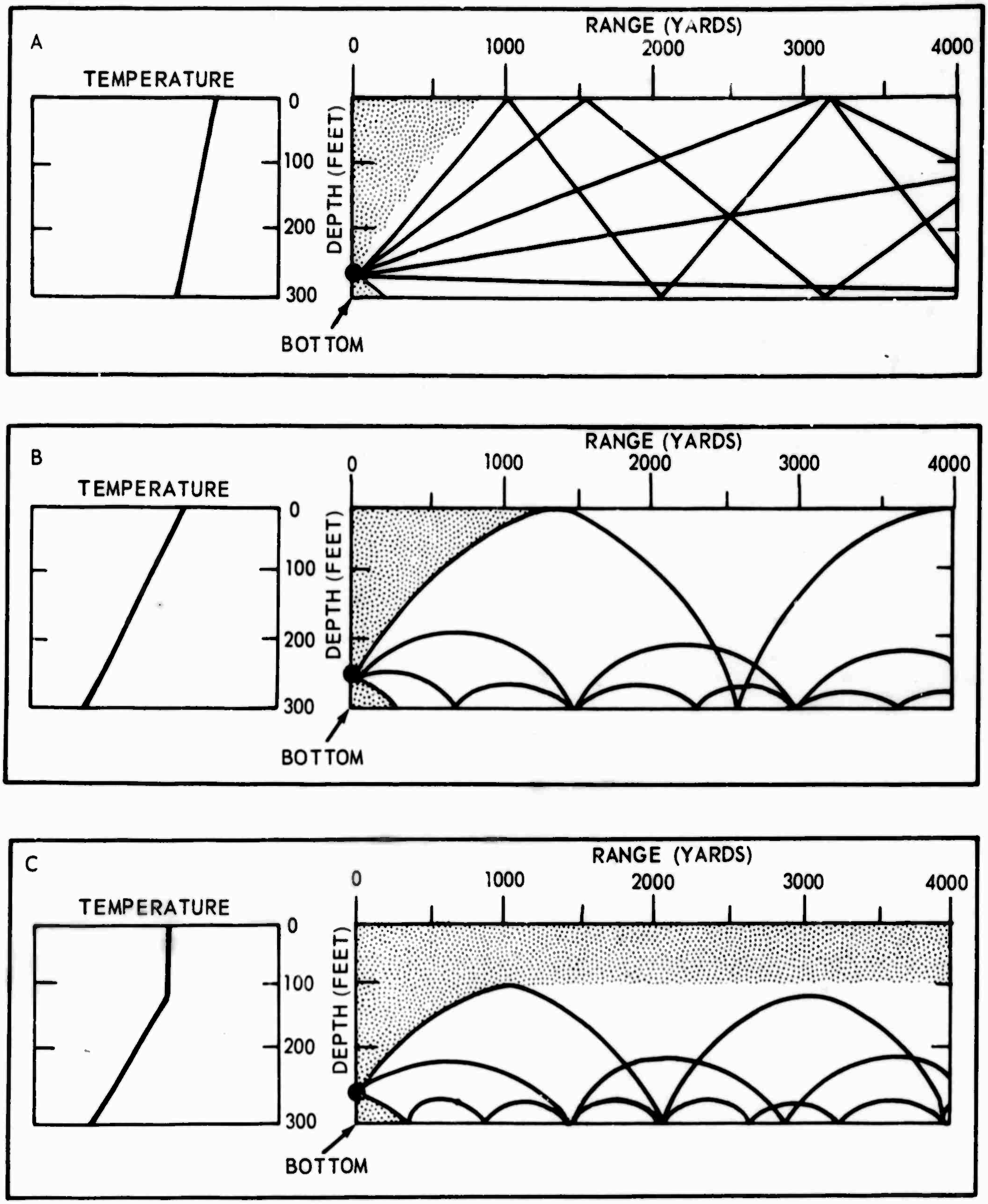


FIGURE 47. EXAMPLES OF SAND BOTTOM REFLECTION

part of the beam travels faster than the lower part, and the beam is diverted downward, leaving a zone near the surface in which direct sound does not enter (shadow zone), except for a weak signal due to scattering and diffraction. If the sound speed increases with depth, the lower part moves faster, and the beam is refracted upward toward the surface. Refraction may bend the sound beam away from a target that might otherwise be detected by an unrefracted beam as shown in Figure 49.

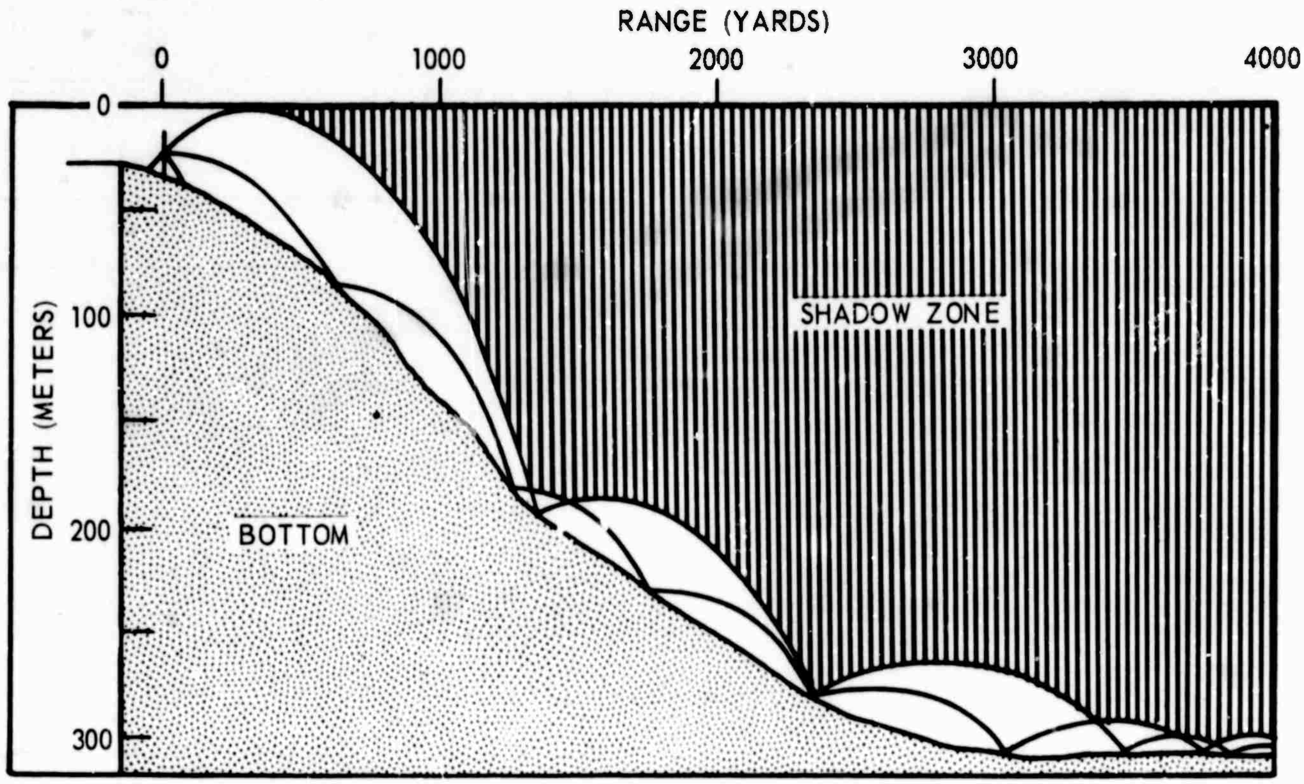


FIGURE 48. EFFECT OF SLOPING BOTTOM ON SOUND REFRACTION

With typical distribution of sound speed with depth, the speed decreases with depth until a minimum is reached at some level below the surface, and below this the speed increases. In Figure 38 the minimum sound speed occurs at about 750 meters. In the tropics this level of minimum sound speed (called the sound channel) may be as deep as 1,800 meters, and in polar regions it may be at the surface. Sound produced at any level tends to be refracted to the level of minimum speed and to remain there, for as it begins to leave this level, it is refracted back toward it, as shown in Figure 50. This, of course, does not refer to sound traveling vertically. If a sound is introduced in the sound channel, as by the explosion of a bomb or depth change, the sound waves start to move outward as expanding spheres, but most of the rays are refracted back toward the minimum sound speed level. Because of this effect, such a sound may travel great distances with relatively little decrease in intensity. Listening gear placed at this

level has detected sounds produced thousands of miles away. This is the principle used in SOFAR (SOund Fixing And Ranging).

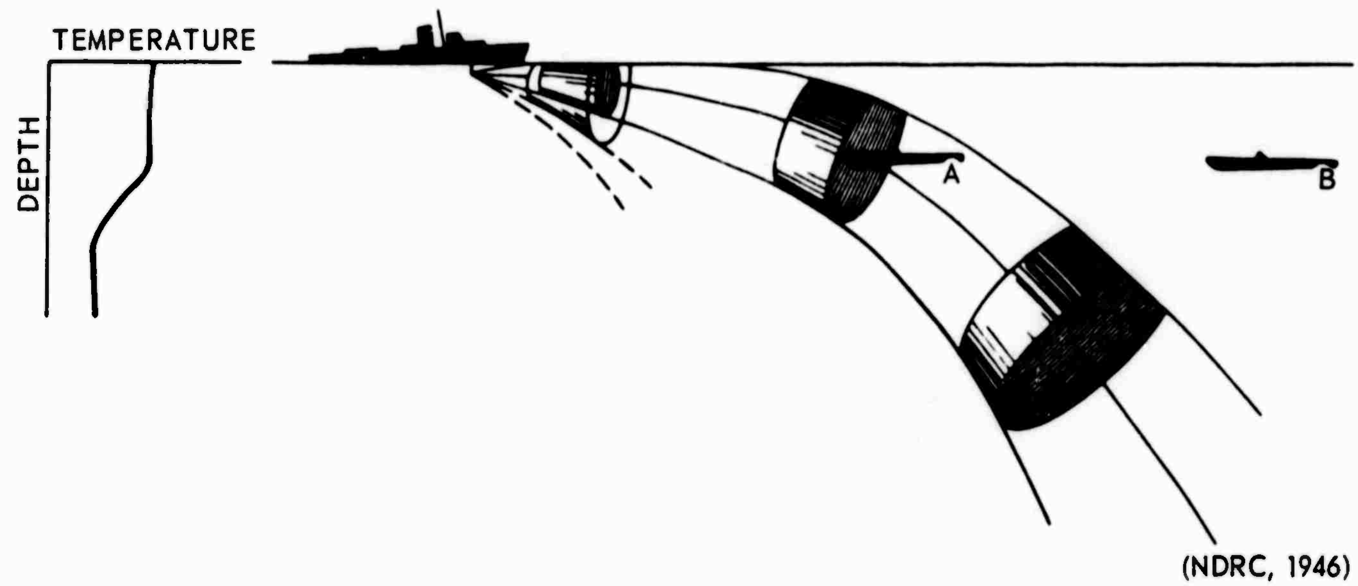


FIGURE 49. EXAMPLE OF REFRACTION OF SOUND BEAM

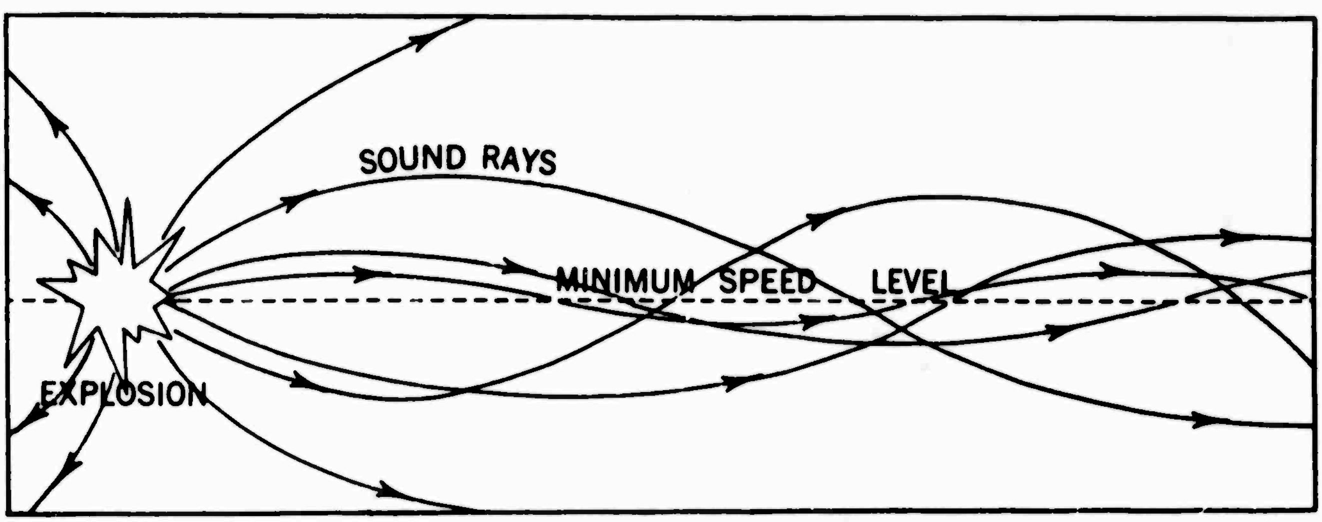


FIGURE 50. TRANSMISSION OF SOUND RAYS ALONG THE MINIMUM SOUND LEVEL

ACOUSTIC RAY THEORY

Sound passing through layers of water having different temperatures and salinities acquires varying speeds resulting in the sound beam being refracted toward lower speeds. When the sound speed structure is complicated by a number of differing gradients, it becomes necessary to construct a diagram of the various paths which a ray or a number of rays may take in

passing from a layer of one constant gradient to a layer with a different constant gradient. The assumption is made that the ocean is stratified, so that the sound speed is the same at all points having the same depth. Echoes are assumed to follow the same paths in returning to the source.

The computation involved in determining ray paths is based on Snell's law, wherein the cosine of the angle of a ray at one sound speed bears a constant relationship to the cosine of the angle of that same ray at any other sound speed. For example, by using values from Figure 51, the computation using Snell's law for the 0° (θ_1) ray initiated at the surface follows:

$$\frac{C_1}{\text{Cos } \theta_1} = \frac{C_2}{\text{Cos } \theta_2} = \frac{C_3}{\text{Cos } \theta_3} = \frac{C_4}{\text{Cos } \theta_4}$$

$$\frac{1498}{\text{Cos } \theta_1 = \theta 1} = \frac{1497}{\text{Cos } \theta_2} = \frac{1496}{\text{Cos } \theta_3} = \frac{1495}{\text{Cos } \theta_4}$$

(Angles are relative to the horizontal)

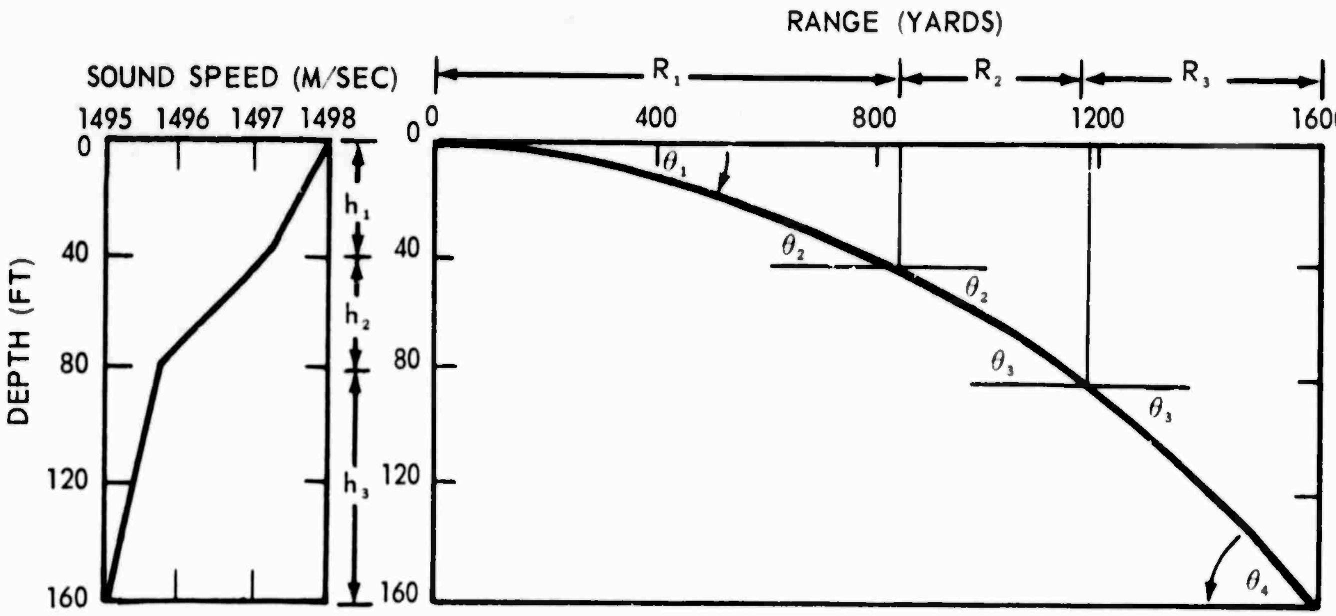


FIGURE 51. PATH OF 0° RAY

Depth (feet) Velocity (M/Sec)		$\frac{\theta_n + \theta_{n\,1}}{2}$	h	cot θ	Range(yds)
0------------1498	$\theta_1 = \theta°$ ---------	--------------	------	------	--------
40-----------1497	θ_2 = 1.83°-------	0.91°--------	h_1 =40	63.66	R_1 = 848
80-----------1496	θ_3 = 3.17°-------	2.50°--------	h_2= 40	22.90	R_2 = 305
160----------1495	θ_4 = 3.67°-------	3.40°--------	h_3= 80	16.75	R_3 = 447

Once the angles at each of the layer interfaces have been determined, the following equations are used to compute the horizontal ranges at discrete depths:

$$h_1 \cot \frac{(\theta_1 + \theta_2)}{2} = R_1$$

$$h_2 \cot \frac{(\theta_2 + \theta_3)}{2} = R_2, \text{ etc.}$$

R_1 is the horizontal range at the 40-foot depth, R_2 is the horizontal range at the 80-foot depth, etc.

The accompanying table gives the computed values for the construction of the ray path illustrated in Figure 51. The angles assigned to rays indicate the true initial angles in degrees. The 0° ray is the horizontal ray as it leaves the source. The part of the beam above this axis is considered to have positive inclination while the part below the axis has negative inclination. Between any 2 ray angles having whole numbers there are an infinite number of ray paths possible. For example, between the -6° and -7° ray there are an infinite number of rays having fractional values such as the -6.32°, -6.44°, and -6.50° rays.

The ray in each layer is a straight line; but if the layers are imagined to become thinner, the ray approaches a smooth curve. Ray theory has the following inadequacies:

(1) Does not account for diffraction (secondary source of sound produced by each point along a ray path) which increases with increasing wavelength.
(2) Does not account for microthermal variations in the ocean.
(3) Does not account for horizontal gradients in the ocean.

ACOUSTIC PROPERTIES OF THE OCEAN BOTTOM

The acoustic properties of the bottom sediments are important in understanding reflection, absorption, and refraction of sound impinging on the ocean bottom, and the effect of these factors on overall transmission loss. In general our knowledge of sound properties on the sea floor is inferred from semi-empirical equations based on correlations of the physical properties of the bottom with sound speed through the bottom. These relationships are of particular importance as they relate to performance of bottom bounce sonar. Bottom reflection losses are defined (in this publication) as energy lost by transmission into the bottom. Scattering losses at the bottom are considered separately.

Present methods of measurement of sound speed in sediments includes seismic research, laboratory studies, and theoretical approaches. In deep water, only the TRIESTE has measured sound speed directly in the sediments by acoustic probes. The problem is complicated because

sediments maybe gross mixtures rather than a homogeneous mass and are usually layered.

It is known that sound speed gradients do exist in the sediments, especially in the upper 200 or 300 meters. Sound speed increases with depth beginning with a sound speed in the surface sediments that is generally equal to or less than the speed of sound in the water at the interface. Sound speed values in the sediments at the interface range between tbout 1470 and 1790 m/sec with over 1/2 of these values equal to or below the sound speed of water at the interface. The elastic properties of the sediments are important for computing sound speed, but there is no way to measure elasticity directly; however, laboratory analysis of sediment densities are routinely made. From the densities the approximate values of porosity can be found, and the gross relationship between porosity and sound speed in the sediments has been established. Porosity is defined as the ratio of the volume of voids between the grains of a sediment sample to the total volume of the sediment aggregate. Density of the sediments is mass per unit volume and is dependent upon densities of the solid particles, the water in the pore spaces, the gas present, and on the relative amounts of these constituents.

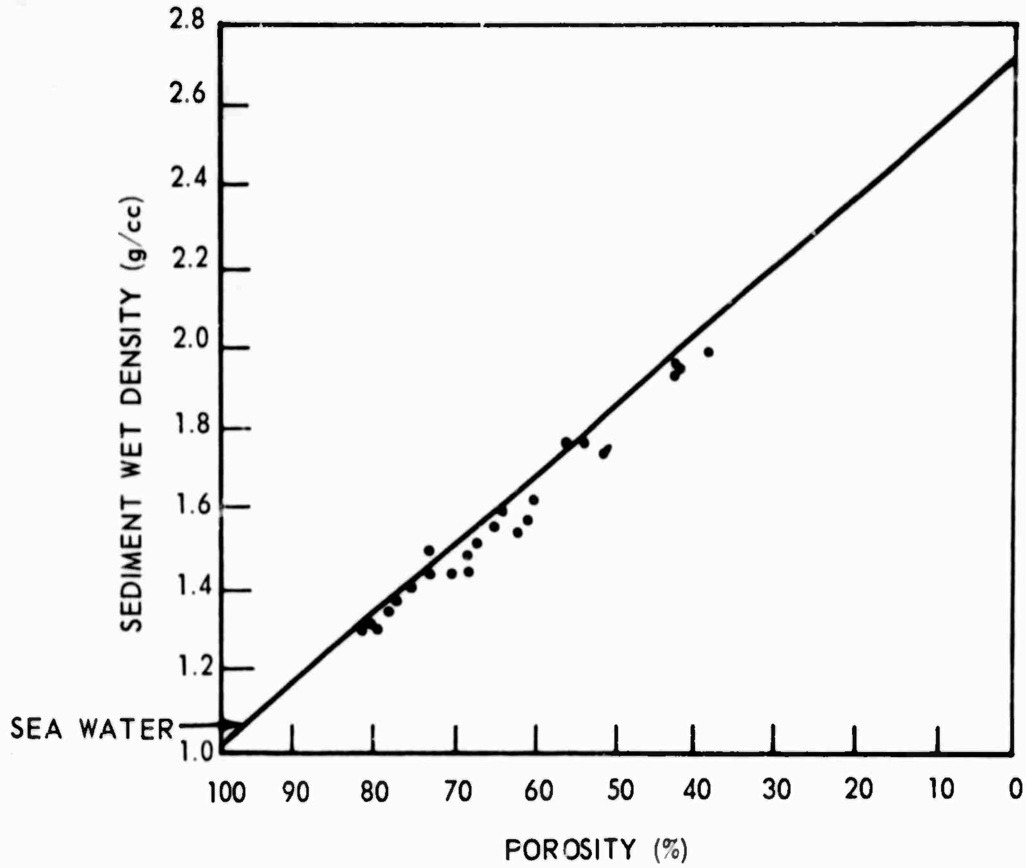


FIGURE 52. SEDIMENT WET DENSITY VERSUS POROSITY

Figure 52 shows the inverse linear relationship existing between density and porosity. Figure 53 shows the relationship of sound speed versus density and porosity. Since density generally increases with depth into the bottom, the bottom scale of Figure 53 may be considered as increasing in depth from left to right with no definite values assigned.

When porosity falls below 77 % the sediment assumes elastic properties and rigidity that are not present in the highly porous suspensions, and sound speeds in the sediments increase rapidly. The specific nature of the sediment particles will to some extent determine the sound speed at a given porosity, but it is believed that porosity is the major influencing factor on compressional sound speed in sediments. Since sound speed gradients do exist in the bottom, then the sound beam will be refracted in the bottom. Present theories also suggest a convergence zone path through the sediments. Listed below are some of the correlations between sound speed and the physical properties of sediments:

(1) Sound speed increases as porosity decreases below 77 %.
(2) Sound speed increases as density increases.
(3) Sound speed increases as grain size increases
(4) Sound speed increases as carbonate content increases.
(5) Sound speed decreases as gas bubbles increase. (A concentration of gas in the sediment of only 1 % causes the same effect on sound speed as a nearly fully gas-saturated condition.)
(6) Increase in pressure contributes little to an increase in sound speed. (Less than 3 % increase in sound speed occurred for an applied pressure change of 400 bars.)
(7) Temperature gradients within the sediments do not appear to be an important factor in determining sound speed; however, computations should start with the temperature of the water at the interface plus a positive gradient of about 0.05°C/m in the first 100 meters of the bottom.
(8) Sound speed increases slightly as the frequency of the sound increases; however, the speed increase will be less than 1 % between frequencies of 20 cps and 1 megacycle.

In determining reflection or transmission into the bottom at the water-bottom interface when sound impinges on the bottom, account must be taken of the differences in acoustic impedance ρc (density times sound speed) at the interface, the grazing angle or the angle that the sound ray makes with the bottom (Fig. 54), and the frequency.

Changes in amplitude and phase of sound energy striking the bottom are described by the Rayleigh reflection coefficient for plane waves. If the speed of sound in the sediment is less than the speed of sound in the water at the interface, then the reflection coefficient decreases from a finite value at normal incidence ($\theta_1 = 0°$) to zero at the angle of intromission (the angle at which all of the sound energy is transmitted into the bottom and none is reflected). When the angle of incidence is greater than the angle of intromission, the reflected wave is reversed in phase by 180°, and the reflection coefficient

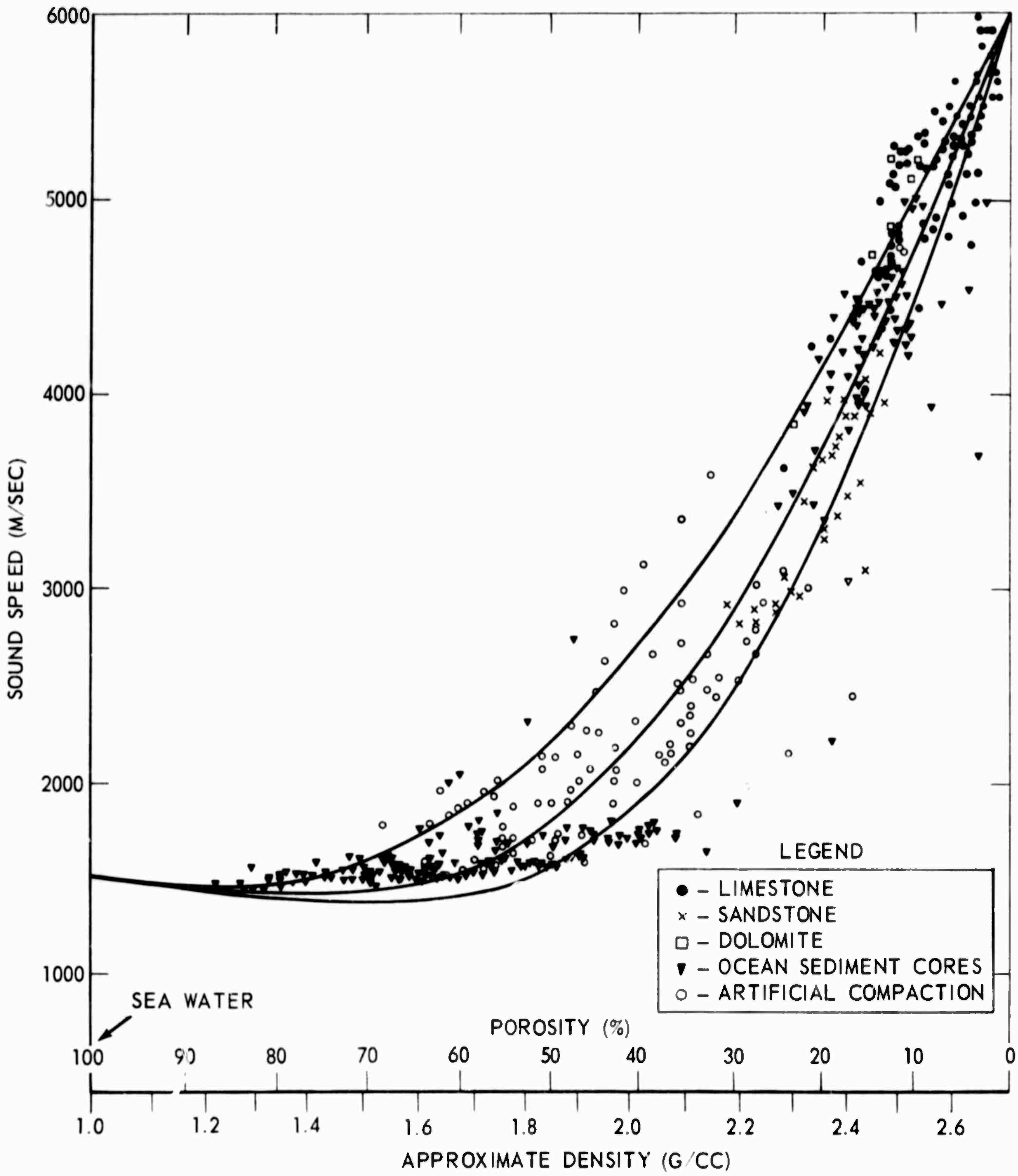


FIGURE 53. COMPRESSIONAL SOUND SPEED-POROSITY DATA FOR SEDIMENTS COMPARED TO EMPIRICAL EQUATIONS OF NAFE AND DRAKE

increases rapidly to a value of one at $\theta_i = 90°$. The angle of intromission (θ_I) may be computed as follows:

$$\theta_I = \cot^{-1} \left[\frac{1 - \left(\frac{c_1}{c_2}\right)^{-2}}{\left(\frac{\rho_2 c_2}{\rho_1 c_1}\right)^2 - 1} \right]^{\frac{1}{2}}$$

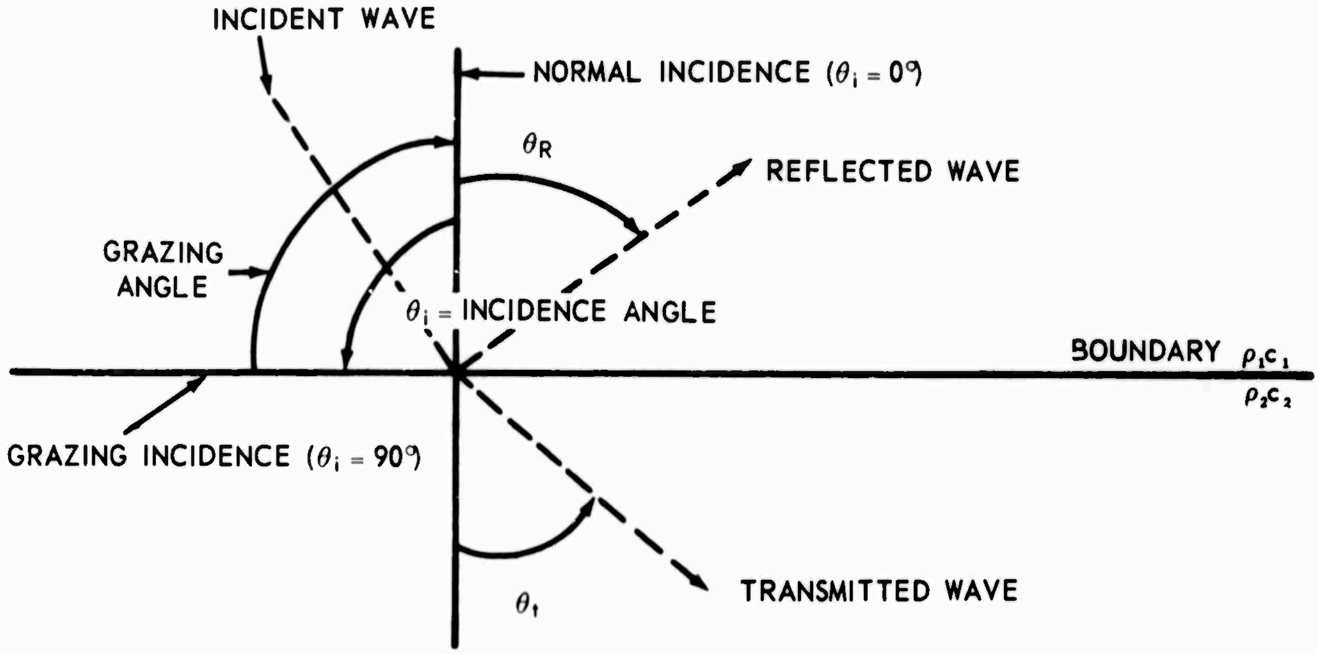


FIGURE 54. TERMINOLOGY AT WATER-SEDIMENT INTERFACE

If the speed of sound in the sediment is greater than the speed of sound in the water at the interface, then the amplitude of the reflected wave will increase from a finite value of less than one at normal incidence ($\theta_i = 0°$) to unity (perfect reflection) at the critical angle and remain unity out to the grazing incidence ($\theta_i = 90°$). The critical angle is defined as $\theta_c = \sin^{-1} \frac{c_1}{c_2}$ where c_1 is the sound speed of the water and c_2 is the sound speed of the sediments at their interface. The amplitude is decreased between angles of normal incidence ($\theta_i = 0°$) and the critical angle because some of the energy is reflected and some is transmitted into the bottom. The phase change is from 0° at the critical angle to 180° at grazing incidence ($\theta_i = 90°$).

When the speed of sound in the sediment (c_2) is greater than the speed of sound in the water (c_1) then the ratio $\frac{c_1}{c_2}$ is less than one and a critical angle will exist; however, if c_2 is less than c_1 the ratio $\frac{c_1}{c_2}$ is greater than one and instead of a critical angle occurring (perfect reflection) there will be an angle of intromission where all energy is transmitted into the bottom. Slight differences in either c_1 or c_2 or both can, thus, result in either perfect reflection or complete loss into the bottom at the appropriate calculated angles. The above generalizations are based mainly on theory. In actuality the situation is more complex, and a critical angle will exist only if the bottom is nonabsorbing. If the bottom is an absorbing medium, then perfect reflection does not occur at the calculated critical angle; instead low losses will occur out to grazing incidence.

The reflection coefficient (R) is computed knowing the density ρ, sound speed c of the sediment and of the water at the interface, and the angle of incidence. The expression is given by:

$$R = \frac{\frac{\rho_2}{\rho_1} - \frac{\sqrt{\frac{c_1^2}{c_2^2} - \sin^2\theta_1}}{\sqrt{1 - \sin^2\theta_i}}}{\frac{\rho_2}{\rho_1} + \frac{\sqrt{\frac{c_1^2}{c_2^2} - \sin^2\theta_1}}{\sqrt{1 - \sin^2\theta_i}}}$$

The loss per reflection may be computed from: Loss in db = 20 log R. When $C_1 = C_2$ the reflection coefficient reduces to

$$R = \frac{\rho_2 - \rho_1}{\rho_2 + \rho_1} .$$

Under this condition there is no dependence on the angle of incidence. When the ratio of the acoustic impedance of the sediment and the water

$$\frac{\rho_2 c_2}{\rho_1 c_1}$$

is greater than one there is an impedance mismatch at the interface, the bottom is considered acoustically hard, and reflection will be good (losses low). The small amount of acoustic energy that does leak through the interface into the bottom will be lost by absorption in the bottom. When the ratio $\frac{\rho_2 c_2}{\rho_1 c_1}$ is less than one there is little impedance mismatch at the interface, the bottom is considered acoustically soft (acoustically transparent), and reflection will be poor (losses high).

Figure 55 illustrates the relationship between porosity and bottom reflection loss and shows a large increase in loss with an increase in porosity. This relationship is understandable since acoustic

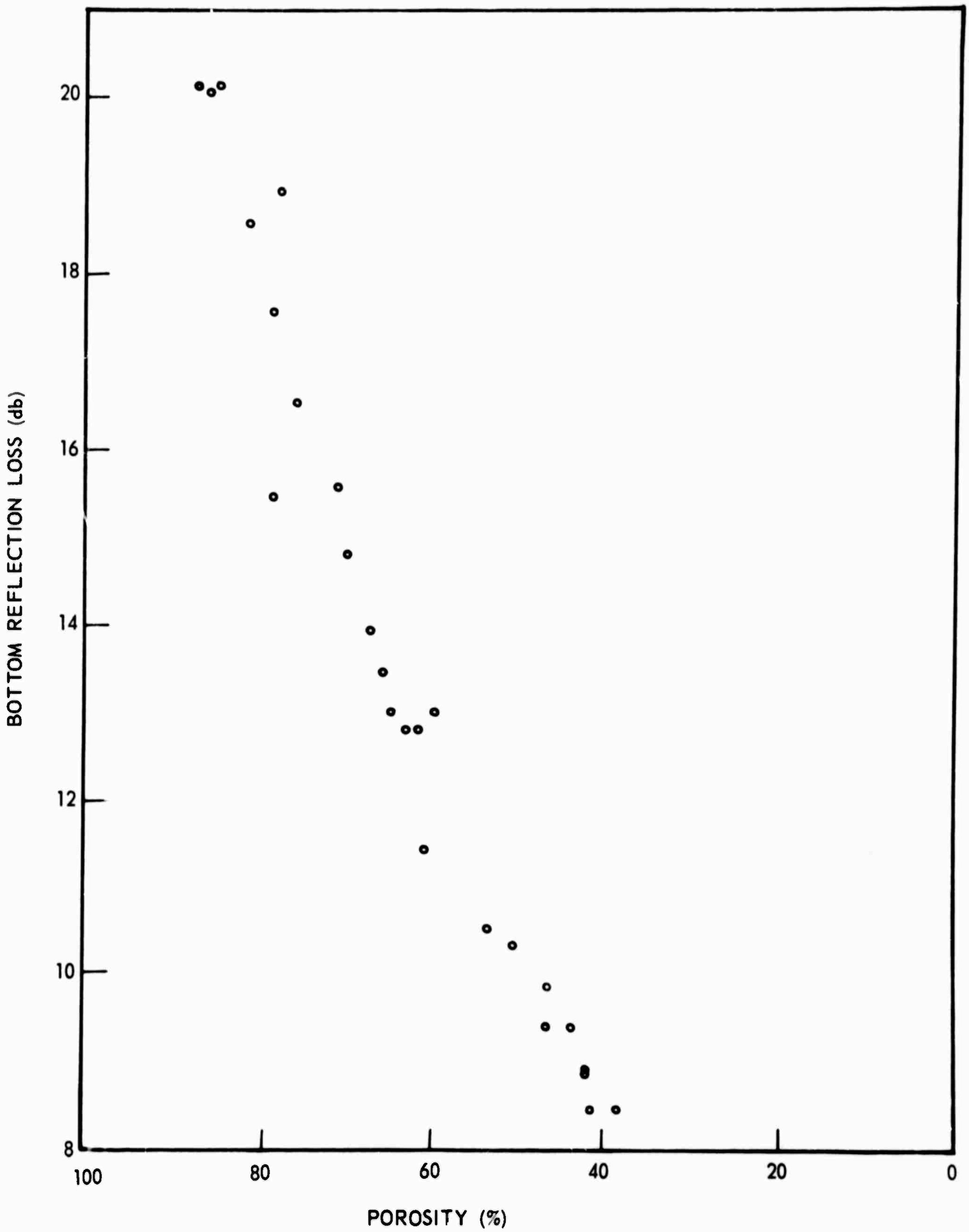


FIGURE 55. COMPUTED BOTTOM REFLECTION LOSS (DB) AT NORMAL INCIDENCE VERSUS POROSITY

impedance decreases as porosity increases, and bottom reflection loss increases with a decrease in acoustic impedance as shown in Figure 56.

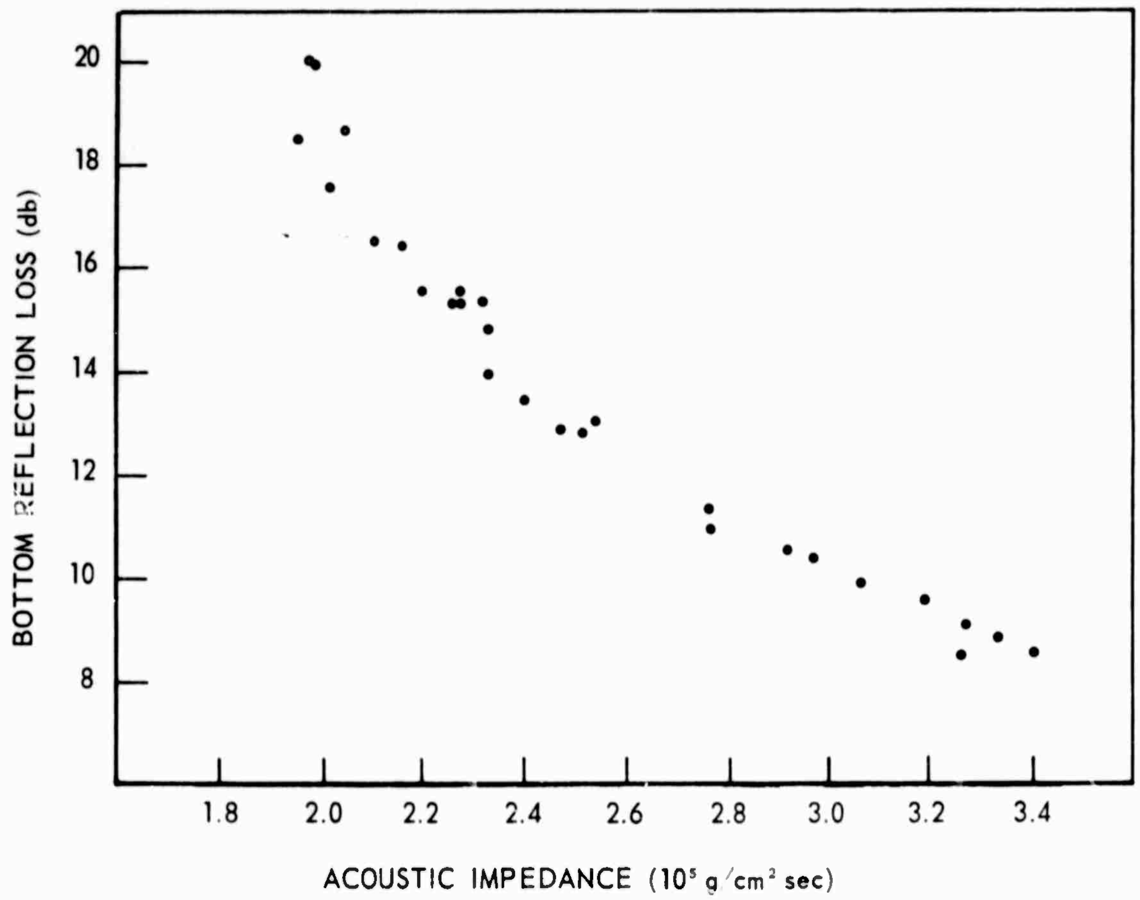


FIGURE 56. BOTTOM REFLECTION LOSS AT NORMAL INCIDENCE VERSUS ACOUSTIC IMPEDANCE

These fundamental relationships and a knowledge of the physical properties at the interface are of value in suggesting the order of magnitude of losses under various conditions. Figures 57 and 58 show the magnitude of loss with changing grazing angles for 2 sets of conditions. Figure 57 (c_1 greater than c_2) illustrates an increase in bottom reflection loss with decrease of grazing angle until all the sound energy is transmitted into the bottom at the angle of intromission. Beyond this point the loss rapidly decreases to zero at normal incidence ($\theta_1 = 0°$).

Figure 58 (c_1 less than c_2) shows a decrease in bottom reflection loss with a decrease of grazing angle out to the critical angle, beyond which all sound energy is reflected and none is lost. This situation is analogous to an imperfect wave guide wherein reflection is efficient only for angles more grazing than critical while for

angles less grazing than critical energy leaks into the bottom. Some researchers have discovered a true wave guide in the surface sediments as evidenced by arrivals of selective frequencies of sound being propagated long distances with low losses.

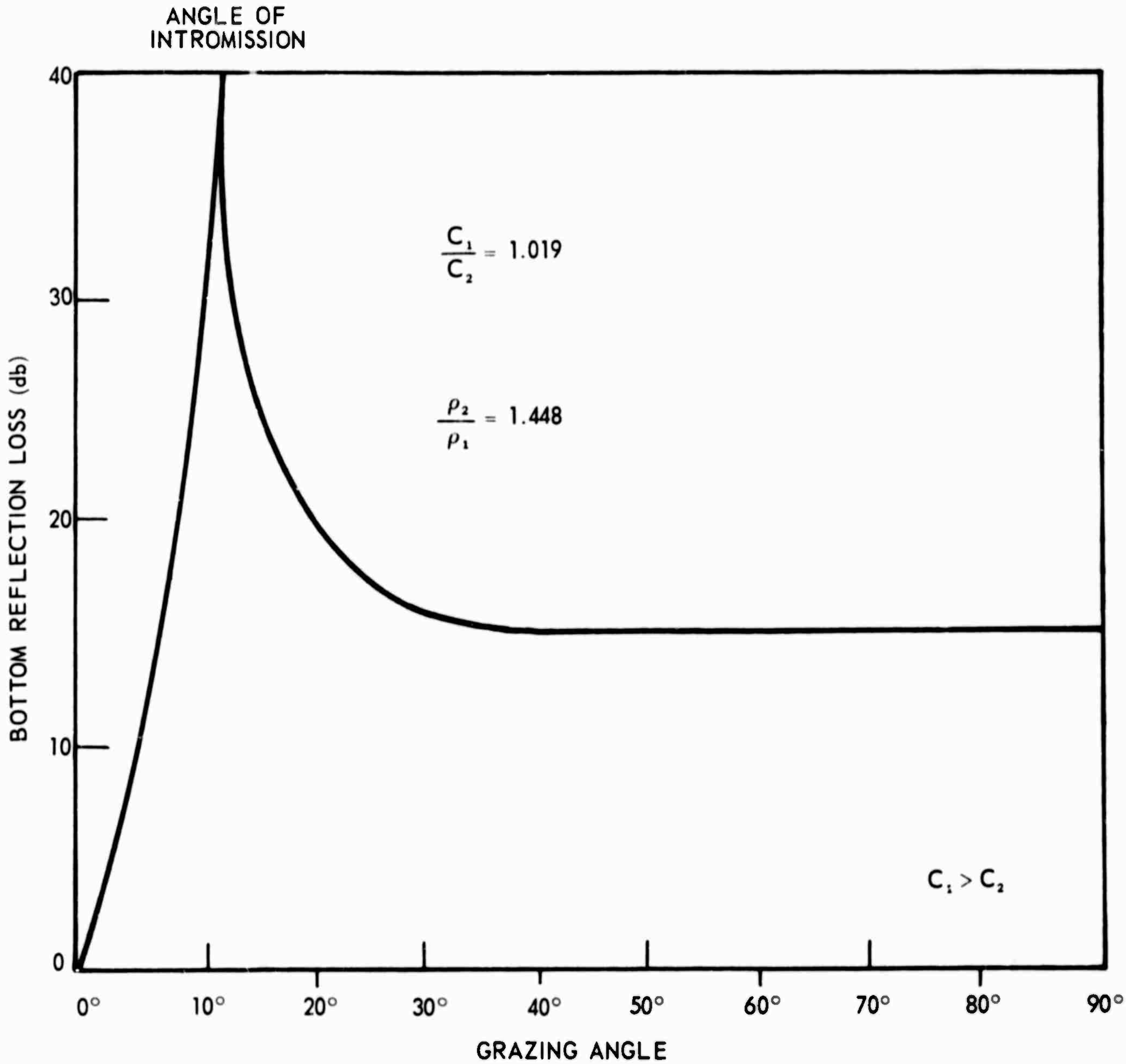


FIGURE 57. BOTTOM REFLECTION LOSS VERSUS GRAZING ANGLE ($C_1 > C_2$)

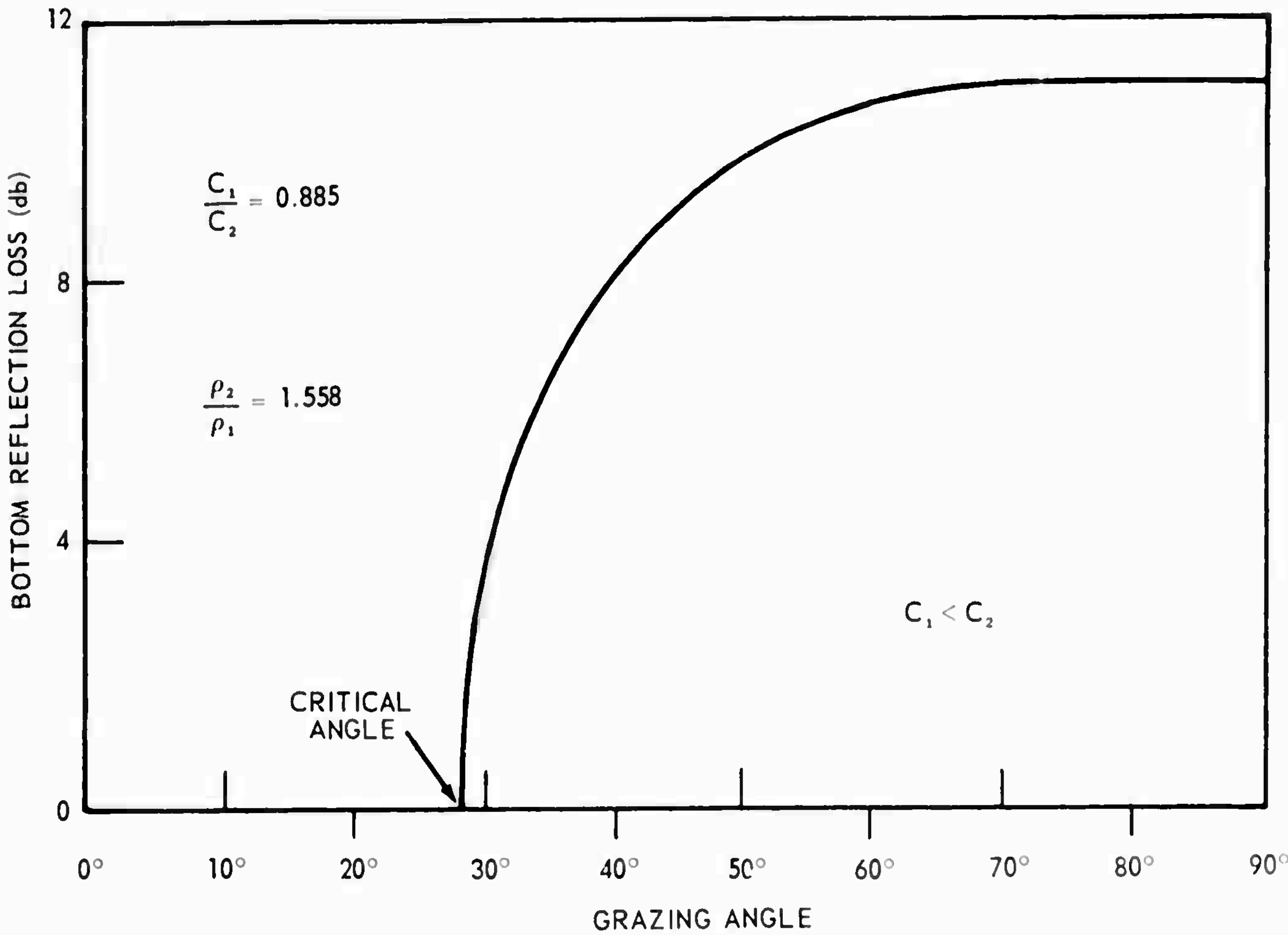


FIGURE 58. BOTTOM REFLECTION LOSS VS GRAZING ANGLE ($c_1 < c_2$)

Table 4 lists selected sediment samples and associated properties. In general it may be stated that fine and medium sands are good reflectors, the coarse to medium silts are fair reflectors, and the fine silts and clays are poor reflectors. Rock, gravel, and coral are poor reflectors because their rough surface tends to scatter sound rather than specularly reflect it. Figure 59 shows curves of bottom scattering from various data sources. For all types of bottoms, scattering is relatively independent of frequency between 1 and 15 kc.

The preceding discussion has dealt with the events occurring at the water-sediment interface; however, the problem is complicated by the layering of sediments with depth and the discovery of sub-bottom reflectors on echo-sounder and seismic records (Figs. 60 and 61). These sub-bottom reflectors are not continuous on the records and may disappear in relatively short distances. The presence of the reflectors appears to indicate layers

TABLE 4 SEDIMENT DATA AND COMPUTED BOTTOM LOSSES

Sediment Type	Median Grain Diam (mm)	Porosity (%)	Wte Density (g/cc)	Sound Speed (m/sec)	Acoustic Impedance (10^5 g/cm^2 sec) (ρ c)	Normal Incidence Bottom Reflection Loss (db)
Medium sand	0.2840	39	1.99	1709	3.40	8.5
Medium sand	0.2700	52	1.50	2039	3.06	9.8
Medium sand	0.2590	42	1.95	1677	3.27	9.0
Medium sand	0.2500	43	1.94	1785	3.46	8.4
Fine sand	0.1760	43	1.96	1699	3.33	8.8
Sandy coarse silt	0.0594	61	1.62	1699	2.76	10.9
Sandy coarse silt	0.0563	57	1.77	1799	3.19	9.5
Sandy coarse silt	0.0540	65	1.60	1496	2.40	13.4
Coarse silt	0.0530	63	1.64	1510	2.47	12.8
Clayey sandy coarse silt	0.0460	66	1.56	1490	2.33	13.9
Medium silt	0.0101	61	1.63	1689	2.76	11.3
Fine silt	0.0094	65	1.60	1590	2.54	13.0
Fine silt	0.0078	69	1.54	1510	2.33	14.8
Fine silt	0.0071	61	1.62	1569	2.54	13.0
Fine silt	0.0070	69	1.44	1600	2.28	15.3
Very fine silt	0.0066	79	1.35	1489	2.01	17.6
Very fine silt	0.0063	52	1.74	1679	2.92	10.5
Clayey very fine silt	0.0058	76	1.41	1493	2.10	16.5
Clayey very fine silt	0.0052	78	1.37	1490	2.04	18.7
Clayey very fine silt	0.0050	--	1.43	1510	2.16	16.4
Clayey very fine silt	0.0046	62	1.57	1600	2.51	12.8
Coarse clay	0.0038	69	1.49	1520	2.27	15.4
Coarse clay	0.0033	55	1.77	1679	2.97	10.3
Coarse clay	0.0030	74	1.50	1520	2.28	15.5
Coarse clay	0.0028	69	1.52	1510	2.27	15.4
Coarse clay	0.0025	83	1.31	1486	1.95	18.5
Coarse clay	0.0020	81	1.32	1490	1.97	20.0
Coarse clay	0.0020	80	1.30	1520	1.98	20.0
Coarse clay	0.0020	69	1.45	1600	2.32	15.3
Coarse clay	0.0019	82	1.31	1540	2.20	15.5

of increased rigidity and acoustic impedance and are probably layers of coarse material such as sand and silt. In the Pacific Ocean one such reflector has been identified as volcanic ash. The presence of sub-bottom layers can usually be explained by turbidity current transport and subsequent deposition. Figure 62 shows how bottom reflection loss varies with changes in acoustic impedance of both the sub-bottom reflecting layer and the layer above it at normal incidence. Especially significant is the effect upon bottom reflection loss of small changes in the thickness of the layer above the reflector. Minimum loss occurs when this thickness is equal to multiples of 1/2 wavelength, whereas maximum loss occurs at odd multiples of 1/4 wavelength. Multipaths from numerous sub-bottom reflectors will also cause differences in loss values due to interference effects. Figure 63 compares

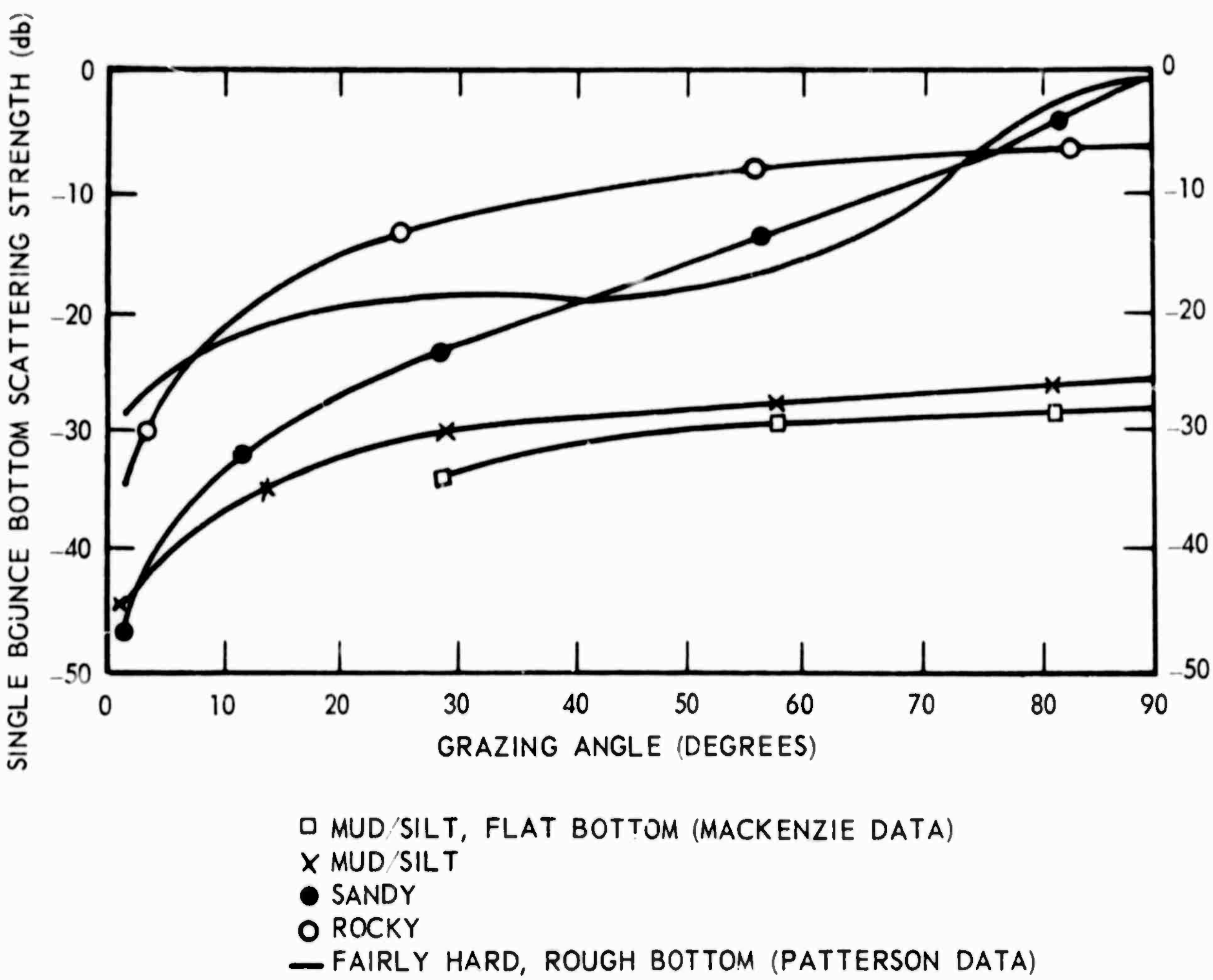


FIGURE 59. EXPERIMENTAL BOTTOM SCATTERING (10 LOG S) DATA

losses obtained in the Pacific Ocean with losses obtained in the Atlantic Ocean. This figure is based on limited data, but does indicate greater losses in the Pacific Ocean which can be explained by the presence of low speed sediments, and because sand carried by turbidity currents is trapped by the deep marginal trenches and is rarely encountered in sediment cores in the central basin. Low speed sediments cause intromission of the sound energy, and the lack of sand layers reduces the number of sub-bottom reflectors; thus, absorption by the bottom occurs.

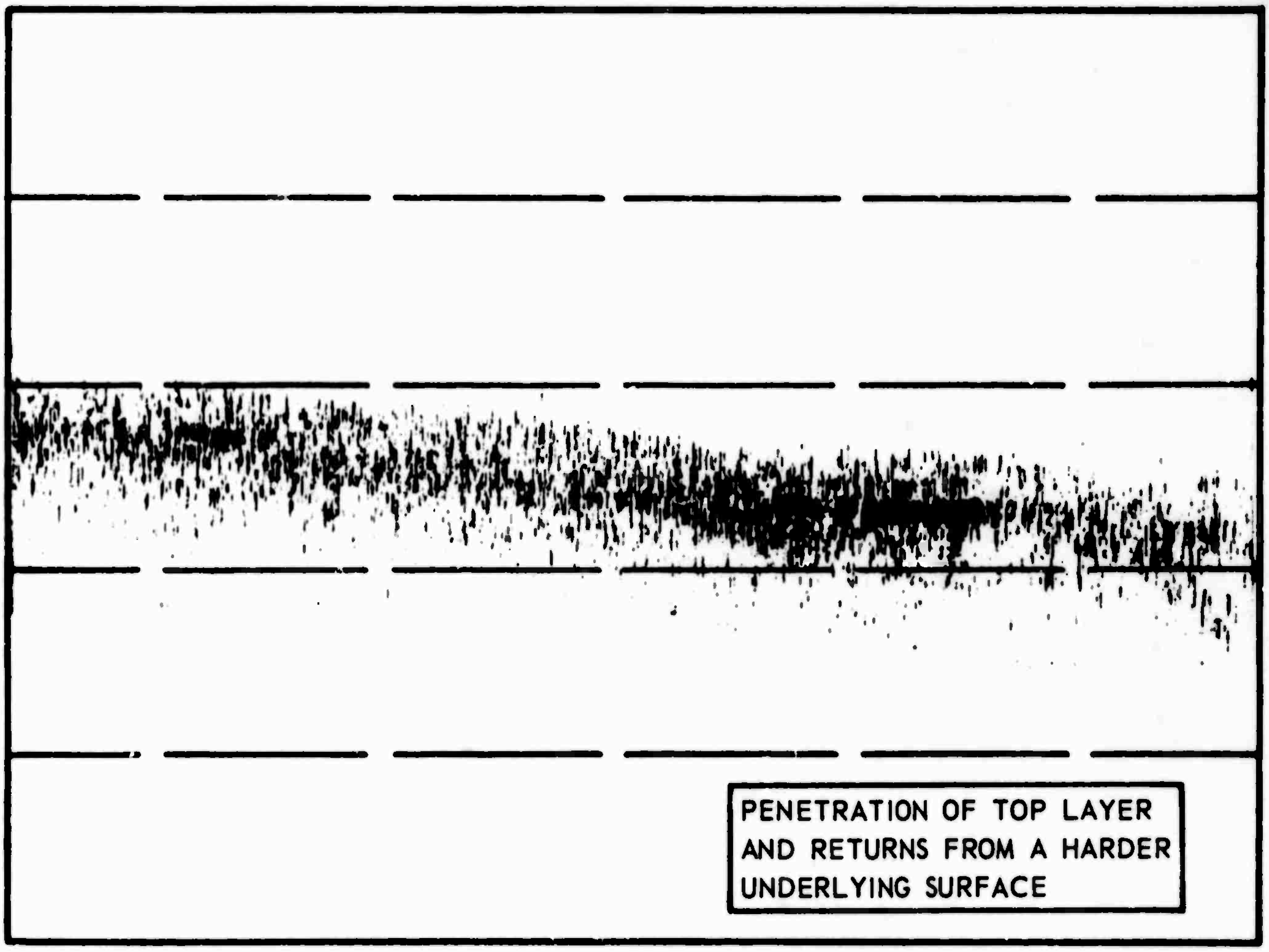


FIGURE 60. SUB-BOTTOM REFLECTOR ON SOUNDING RECORD

Figure 64 illustrates fluctuations in 12-kc bottom reflection loss at normal incidence encountered during a 13-minute period at 2 drift stations in the Atlantic Ocean less than 100 miles apart. It is not known if these fluctuations result solely from variations in the sediments as the ship drifted or to what extent bottom scattering, internal fluctuations in the sound field, and pulse-by-pulse variations in the transmitting equipment affects the recordings. If these variations are caused by sediment changes then it will become necessary to study the microstructure of the sea floor

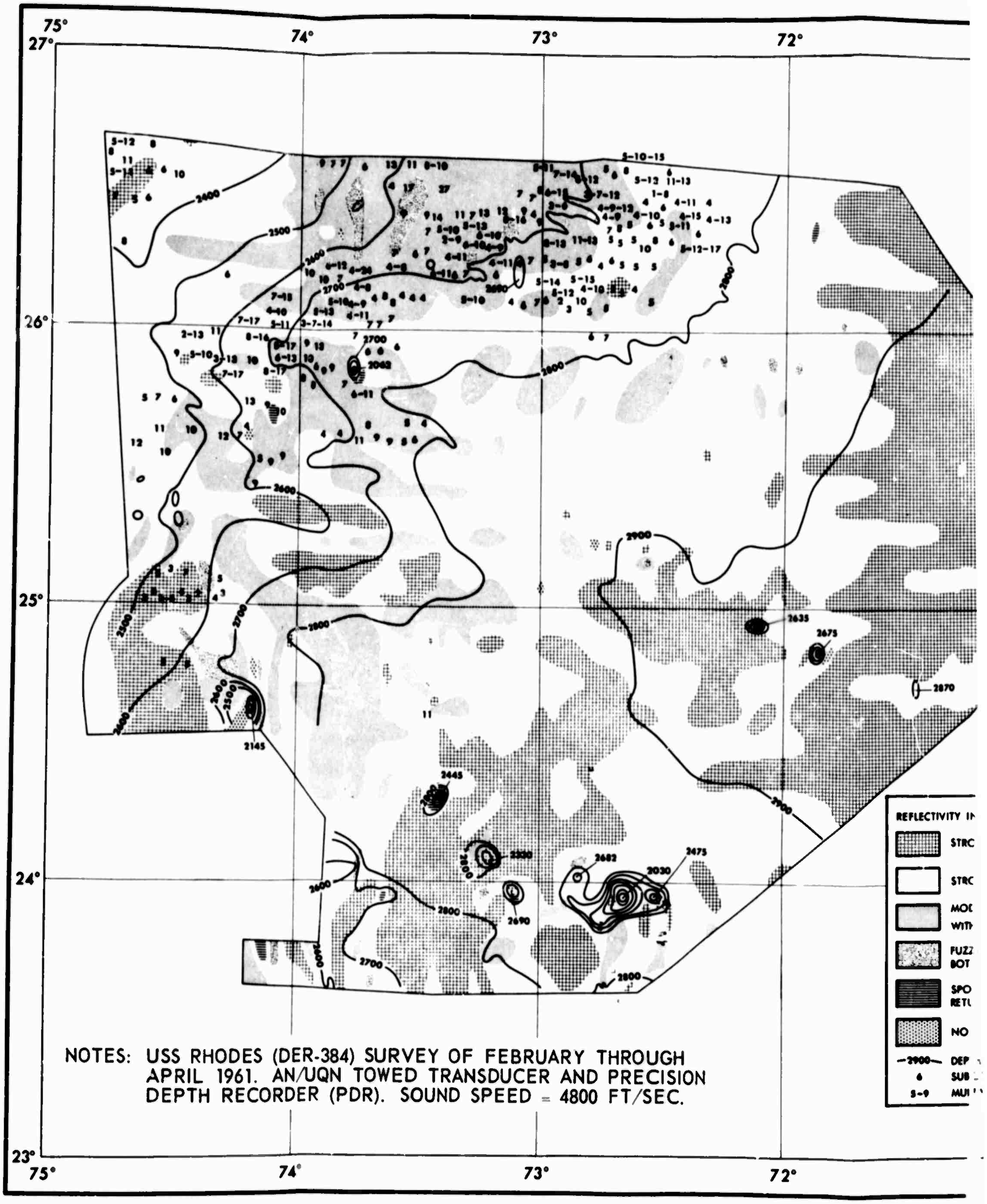
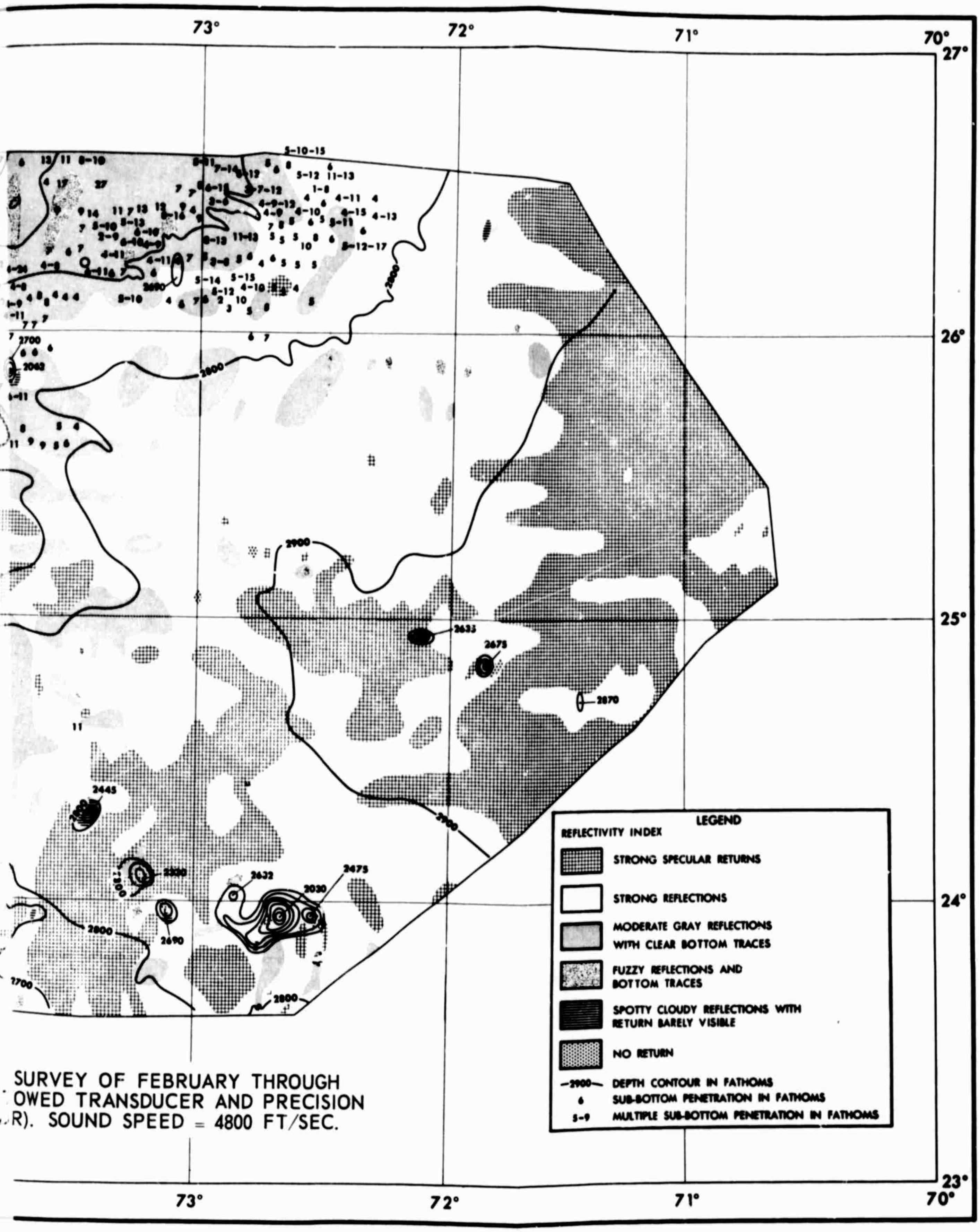


FIGURE 61. REFLECTIVE INDEX AND SUB-BOTTOM PENETRATION VALUES FO
II AREA B

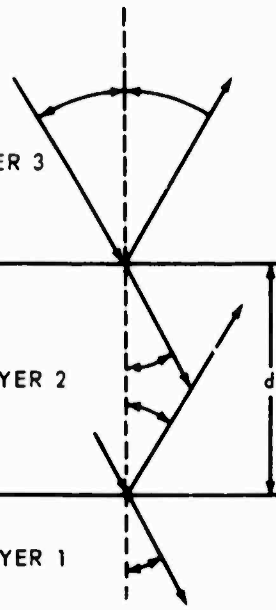
A



REFLECTIVE INDEX AND SUB-BOTTOM PENETRATION VALUES FOR HOLIDAY II AREA B

B

THREE-LAYER BOTTOM REFLECTION MODEL



THREE-LAYER BOTTOM REFLECTION MODEL		CURVE 1	CURVE 2	CURVE 3
		————	– – – – –	....
LAYER 3	WATER	$\rho_c = 1.63$	1.63	1.63
LAYER 2	CLAY OR FINE-GRAINED SILT	$\rho_c = 2.28$	2.20	2.18
LAYER 1	COARSE SILT OR SAND	$\rho_c = 3.06$	2.54	2.92

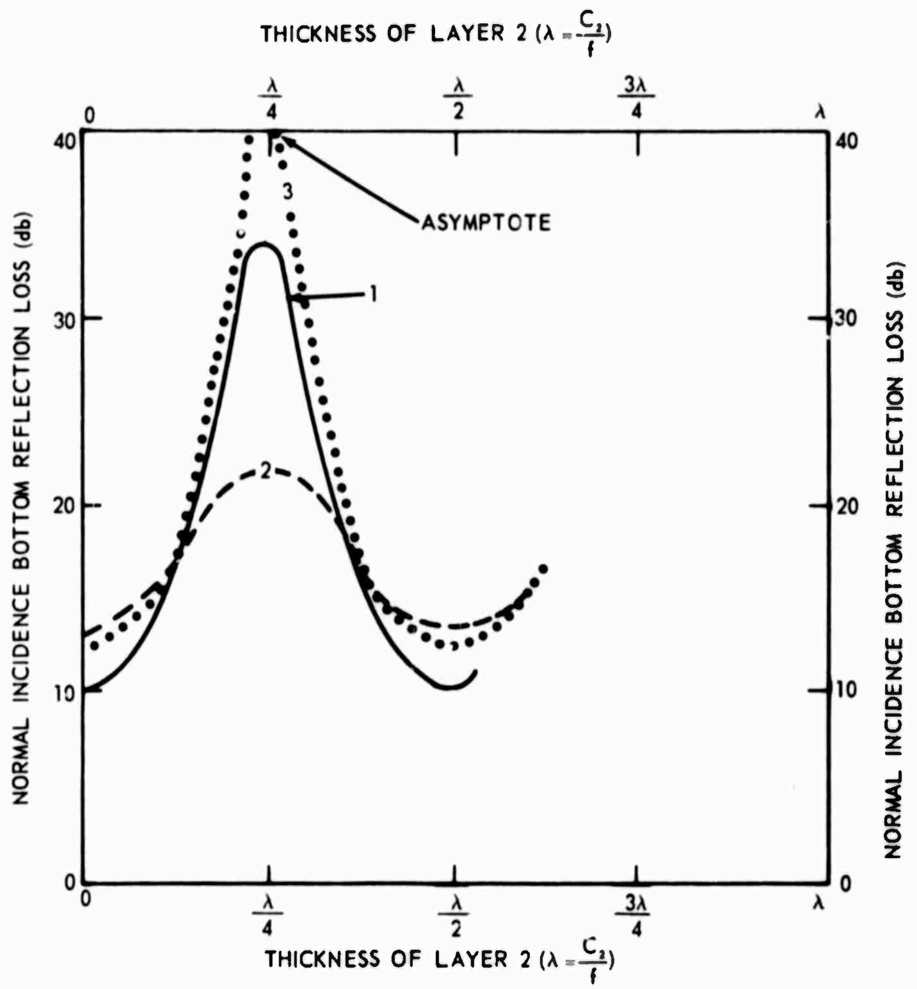


FIGURE 62. BOTTOM REFLECTION LOSS VERSUS THICKNESS OF LOG

for an explanation. The thickness of layer 2 (Fig. 62) also is important. If layer 2 is thick, more variations in bottom reflection loss versus grazing angle occur at a given location. If layer 2 is relatively thin, less variations in bottom reflection loss versus grazing angle occur. Therefore the study of microstructure of the bottom must also include a study of layer 2.

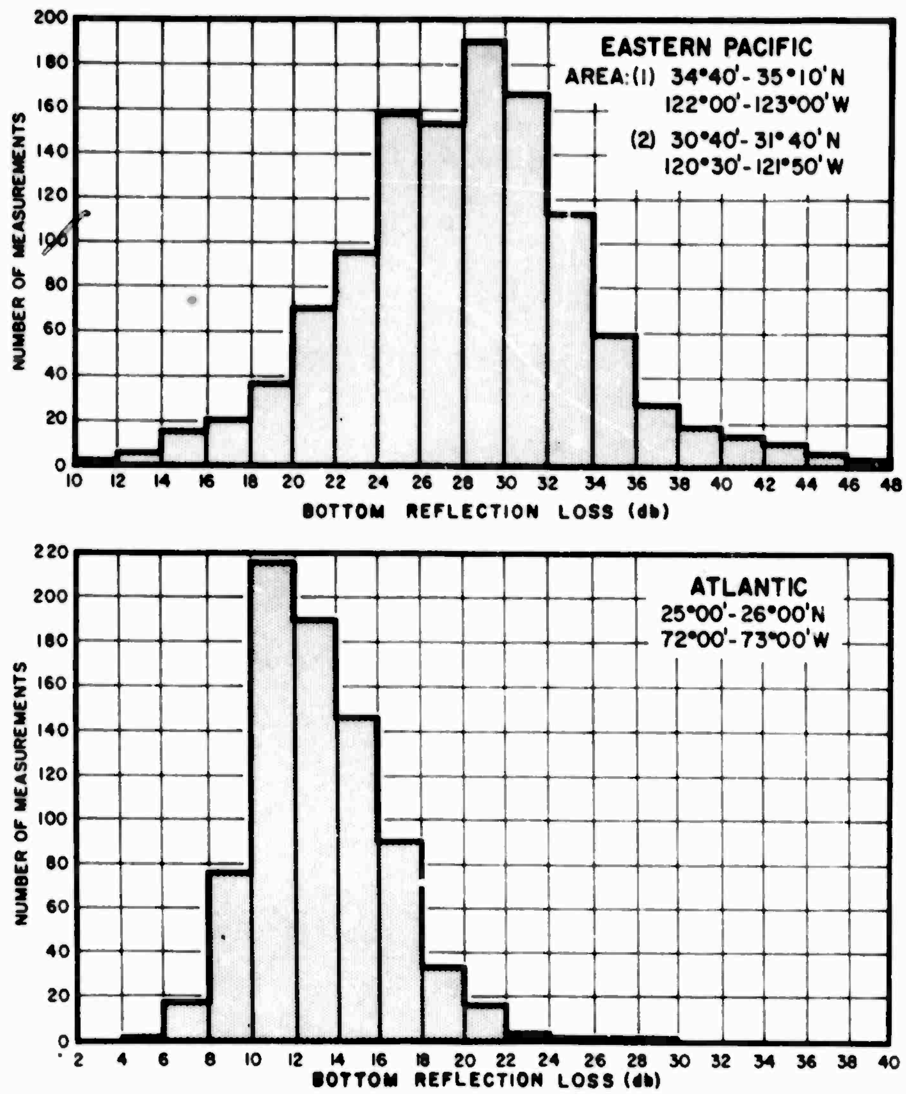


FIGURE 63. FREQUENCY DISTRIBUTION OF BOTTOM REFLECTION LOSS FOR 12-KC NORMAL INCIDENCE SIGNALS

Computations and comparisons of bottom reflection loss data for various parts of the ocean are more meaningful when an absorption coefficient of sound in sediments is included. A combination of reflection, scattering, and absorption loss yields the total loss at the sediment interface. Various studies have related absorption to porosity and to grain size of the sediments. In general absorption, as shown in Figure 65,

is maximum for sediments of intermediate porosity (45 to 60 %) where coarse silt or sand is dominant. Lowest absorption occurs in sediments having high porosity (clays and fine silts).

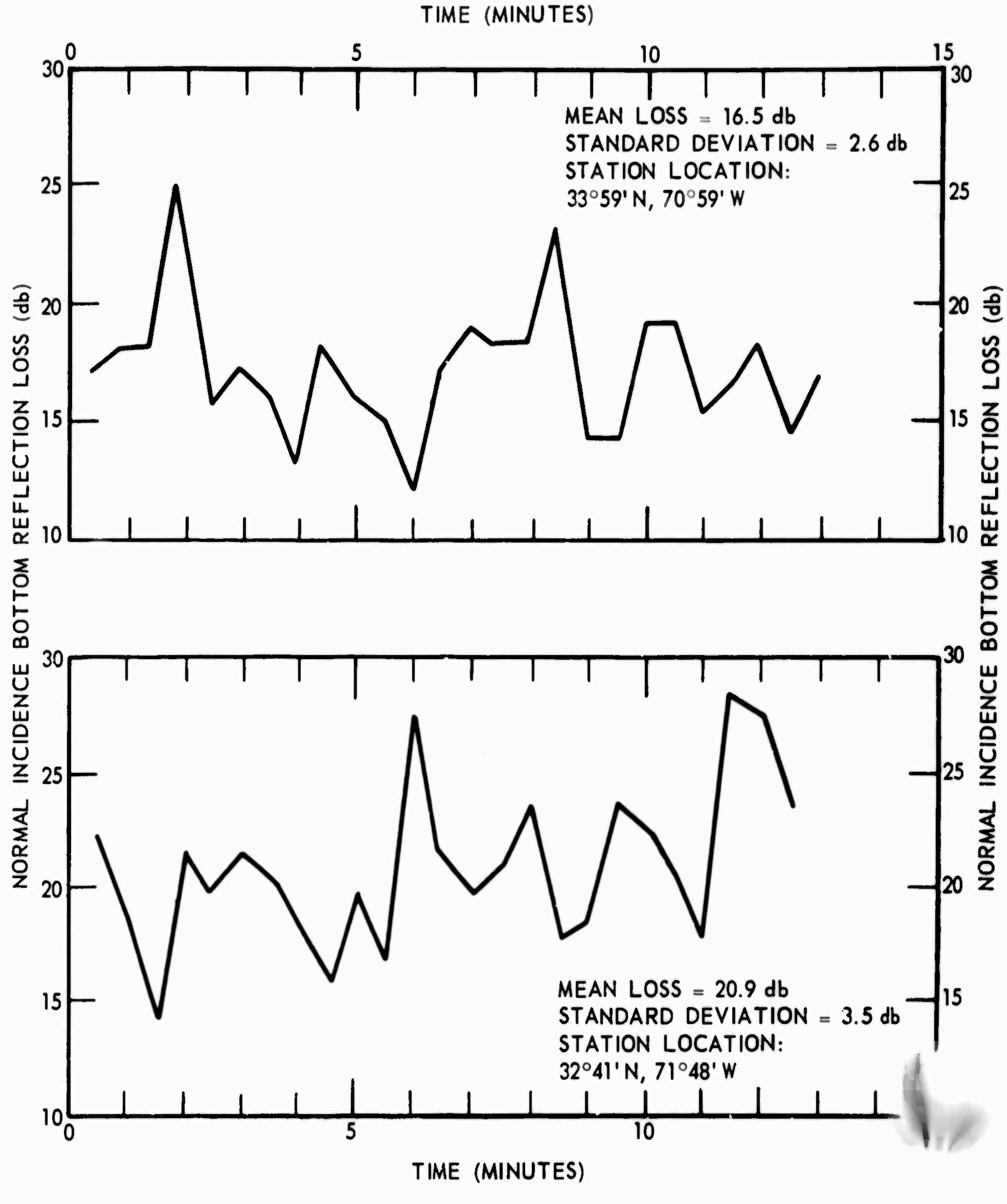


FIGURE 64. FLUCTUATIONS IN BOTTOM REFLECTION LOSS

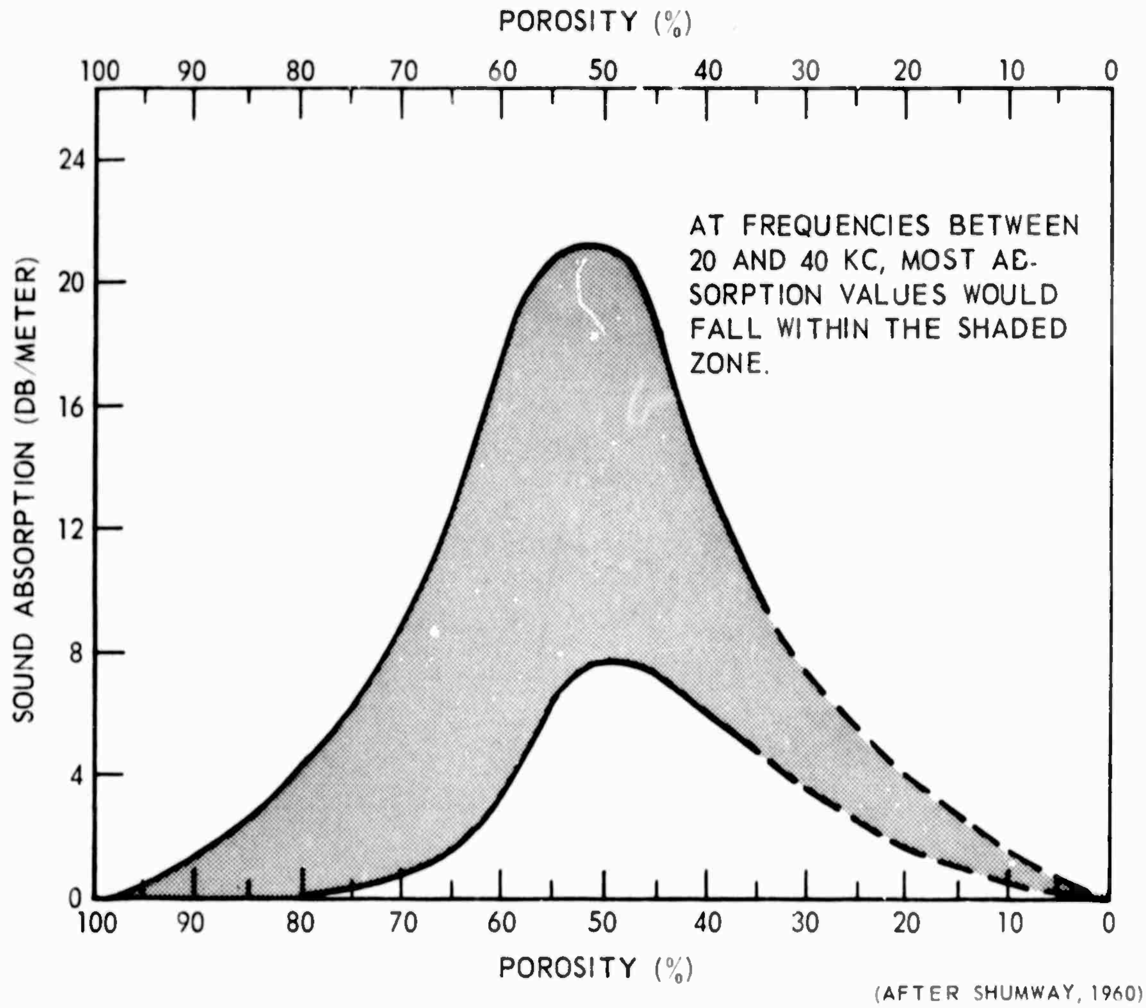


FIGURE 65. SOUND ABSORPTION VERSUS POROSITY FOR UNCONSOLIDATED SEDIMENTS

The absorption coefficient has been shown to be frequency dependent; however, the values will vary between the extreme limits of the square of the frequency (for low frequencies) to the square root of the frequency (for high frequencies). Figure 66 taken from data at 25 stations in a one-degree square 125 miles northeast of the Bahama Islands, shows a definite dependence of frequency as a function of bottom reflection loss at normal incidence. Average losses at 1 kc are about 11 db less than losses at 10 kc. Exact relationships are lacking because of the many additional factors involved such as frictional loss, viscous drag, elastic hysteresis, acoustic surface area of the grain particles, presence of gas bubbles, etc.; however, in general it may be said that at the bottom, low frequency sound will penetrate deeper into the bottom than high frequency sound.

Thin reflective silt layers within about 2 wavelengths of the water-bottom interface are thought to be the source of reduction in bottom loss. At 3.5 kc (wavelength of 1.4 feet) reflections within the first 3 feet of

the bottom contribute to the bottom arrival data. Constructive interference occurs where the thin silt layers are 1/4 wavelength thick (0.35 foot) because the phase arrivals between the top of the thin layer and the bottom of the thin layer are 180° apart and therefore add. Theoretical calculations show that at normal incidence the combination of losses by both scattering and thin silt layers in the first 3 feet of the bottom would be about 10 db. In contrast, the absorption losses in the first few feet of the bottom at 3.5 kc are negligible. Amplitudes of sub-bottom reflectors to depths of 50 feet into the bottom at 3.5 kc can be comparable to the bottom interface reflection. Below 50 feet the amplitudes decrease.

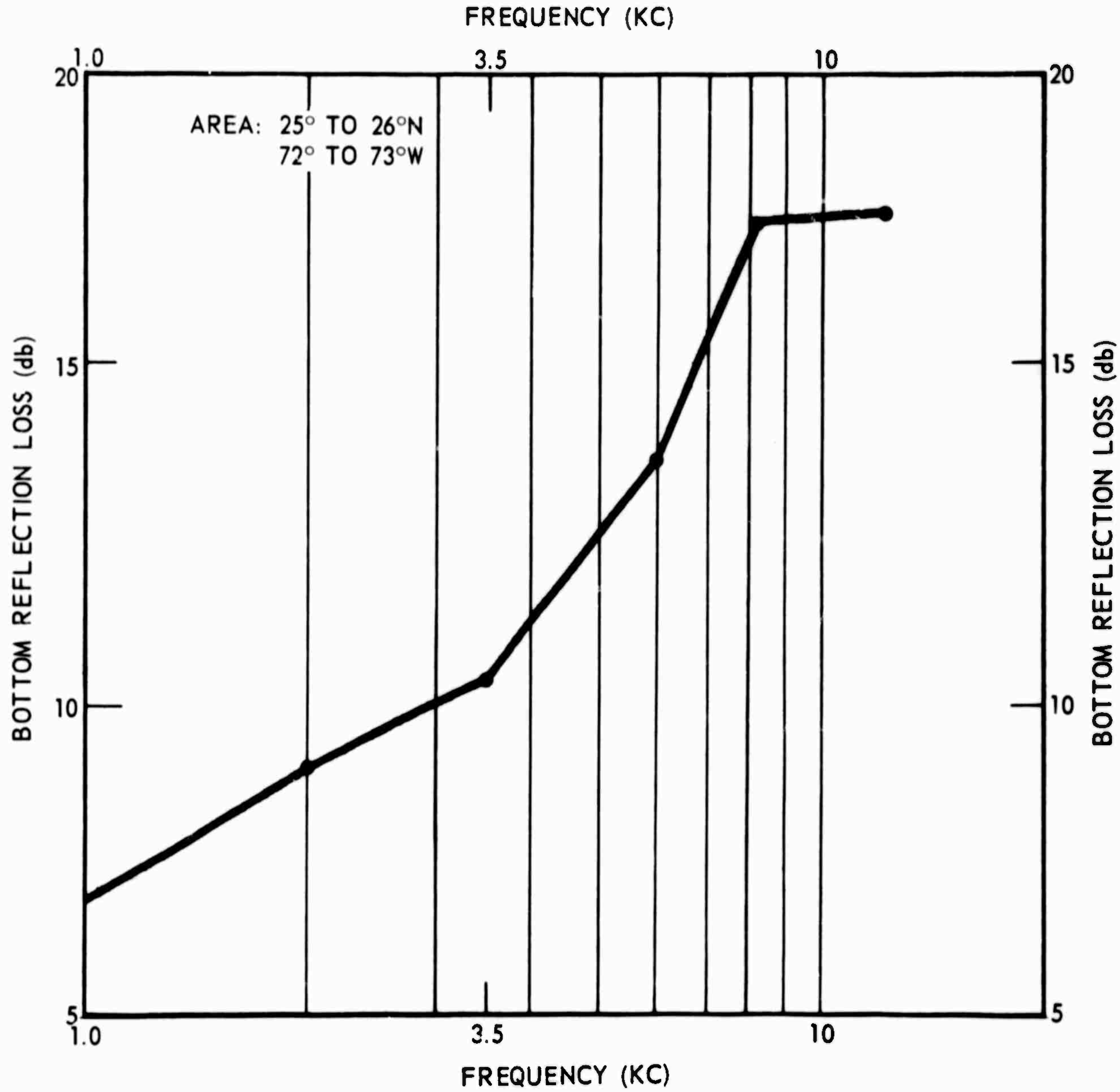


FIGURE 66. AVERAGE BOTTOM REFLECTION LOSS VERSUS FREQUENCY AT NORMAL INCIDENCE FOR 25 STATIONS

BACKGROUND NOISE

In the ocean, background noise is any sound which interferes with the reception of a wanted signal. Unlike reverberation, it is independent of the amount of energy transmitted by the sound source and covers a wide range of frequencies. Background noise usually is considered to consist of self noise and ambient noise. Self noise is that noise associated with the electronic and mechanical operation of the sonar and the platform; ambient noise encompasses all of the natural noise in the sea.

If the marine environment were noisefree, detection of an acoustic signal would still be difficult because of the noise inherent in the sound equipment, in the platform on which it is mounted, and in the motion of the platform. Even when the sound gear is towed separately, noise is generated by water movement around the unit and supporting cable. This noise is known as own-ship's noise or self noise. Self noise may result from: 1) Circuit noise arising from noisy tubes and components, 2) transducer noise caused by water turbulence around the housing, 3) hull noise arising from loose structural parts, or 4) machinery noise from propulsion or auxiliary equipment. Sometimes, self noise is referred to as machinery noise, propeller noise, and hydrodynamic noise. Since self noise is not part of the oceanographic environment, it often can be reduced by various quieting schemes.

A surface vessel produces considerable noise which can interfere with target detection. The sources of noise are extremely diversified and are distributed through the entire range of frequencies. The chief sources and the paths by which they reach the receiving transducer are shown in the following table:

Source	Paths	Character and Importance
MACHINERY NOISE AND VIBRATION (a) Main Propulsion (b) Main Shaft and Bearing (c) Auxiliaries	(i) Via hull to transducer support (ii) Via hull to water	Contains line components. Spectrum falls at about 12 db/octave above 1–2 kc. Important at low frequencies on submarines, surface ships, and ship-towed sonar.
PROPELLER NOISE	(i) Via direct path and reflection within dome. (ii) Reflection and scattering from sea bottom. (iii) Reflection and scattering from sea surface. (iv) Reflection and scattering from other reflectors present in medium and introduced by passage of the vessel. (v) Reflection, scattering, and re-radiation from hull. (vi) Shaft to hull	Continuous spectrum (cavitation noise) falling at 6 db/octave. Level increases rapidly with speed at inception of cavitation. Path (ii) particularly important in surface ships in shallow water. Path (iii) particularly important in torpedoes.
HYDRODYNAMIC NOISE (a) Flow Noise (i) Noise due to the eddies and vortices themselves. (ii) Pressure variations on hydrophone element. (b) Flow Excitation—Vibration induced in transducer (or dome), hull superstructure, or cable support. (c) Cavitation—Around transducer (or dome), support, hull appendages, and imperfections. (d) Bubbles (i) Striking dome (ii) Deformed in pressure field near dome. (iii) Coalescing and breaking up near dome. (e) Surface Waves—Generated by ship's passage.	Direct paths	Continuous spectrum (a, ii), (b) Important in submarine listening at low frequencies. (c) Important on surface ships and torpedoes at high speeds. Importance in surface ships depends on condition of hull or dome. (d) Important for surface ships. Importance depends on dome position and shape of hull. (e) Of doubtful importance.
CIRCUIT NOISE Thermal Noise Tube Noise Hum Microphonics	Generated within system	Contribution negligible in well-designed systems.
MISCELLANEOUS (a) Helicopter Rotor Noise (b) Crew Movement	(a) Direct from air to water (b) As for machinery noise and vibration.	(a) Limiting in helicopter-borne sonar. (b) Important in submarine listening.

The overall levels of noise (frequencies of 0.1 to 10.0 kc) for destroyers at speeds from 10 to 25 knots range from 5 to 40 db. The relative spectrum frequency distribution is shown in Figure 67. The data have been averaged for 52 ships and give the spectrum level minus the overall level.

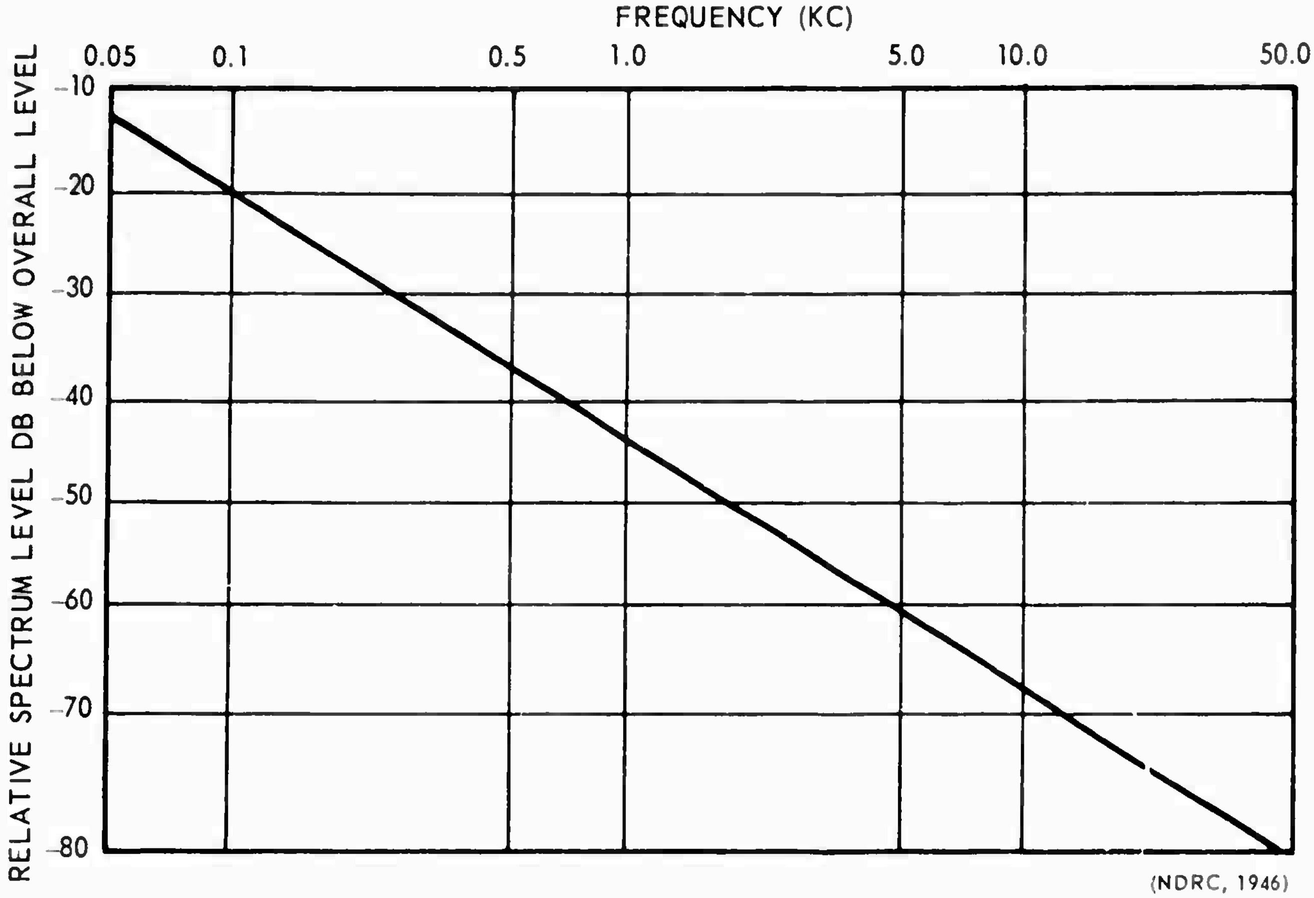


FIGURE 67. RELATIVE SPECTRUM LEVEL VERSUS FREQUENCY FOR SURFACE SHIPS

Machinery noise is produced by the main propulsion plant, reduction gears, propeller shafts, auxiliary machinery, and the underwater discharges from the ship. The sounds include whines, squeaks, or grumbles of various frequencies. This noise is of greatest importance at low speeds as it is then concentrated in the low frequency range.

The sonar gear on submarines when used for listening, hears lower frequencies over a wider range compared to surface ship sonar. The interfering sounds produced are mainly from the same sources as in a surface ship. Machinery noise of a submarine is characterized by line components that stand out above a continuous background at low frequencies, and in contrast to other forms of self noise, is relatively independent of speed. Since the auxiliaries of submarines are of great importance in ultraquiet operations, particular attention has been paid to suppression of these noise sources. The overall level of a few of these noise sources is listed in Figure 68.

These sounds differ in quality and intensity from one submarine to another. In addition, the output of various pieces of equipment changes with time. At low speeds the principal source of noise above 20 kc is ambient noise, but this noise yields to propeller noise at speeds when cavitation occurs.

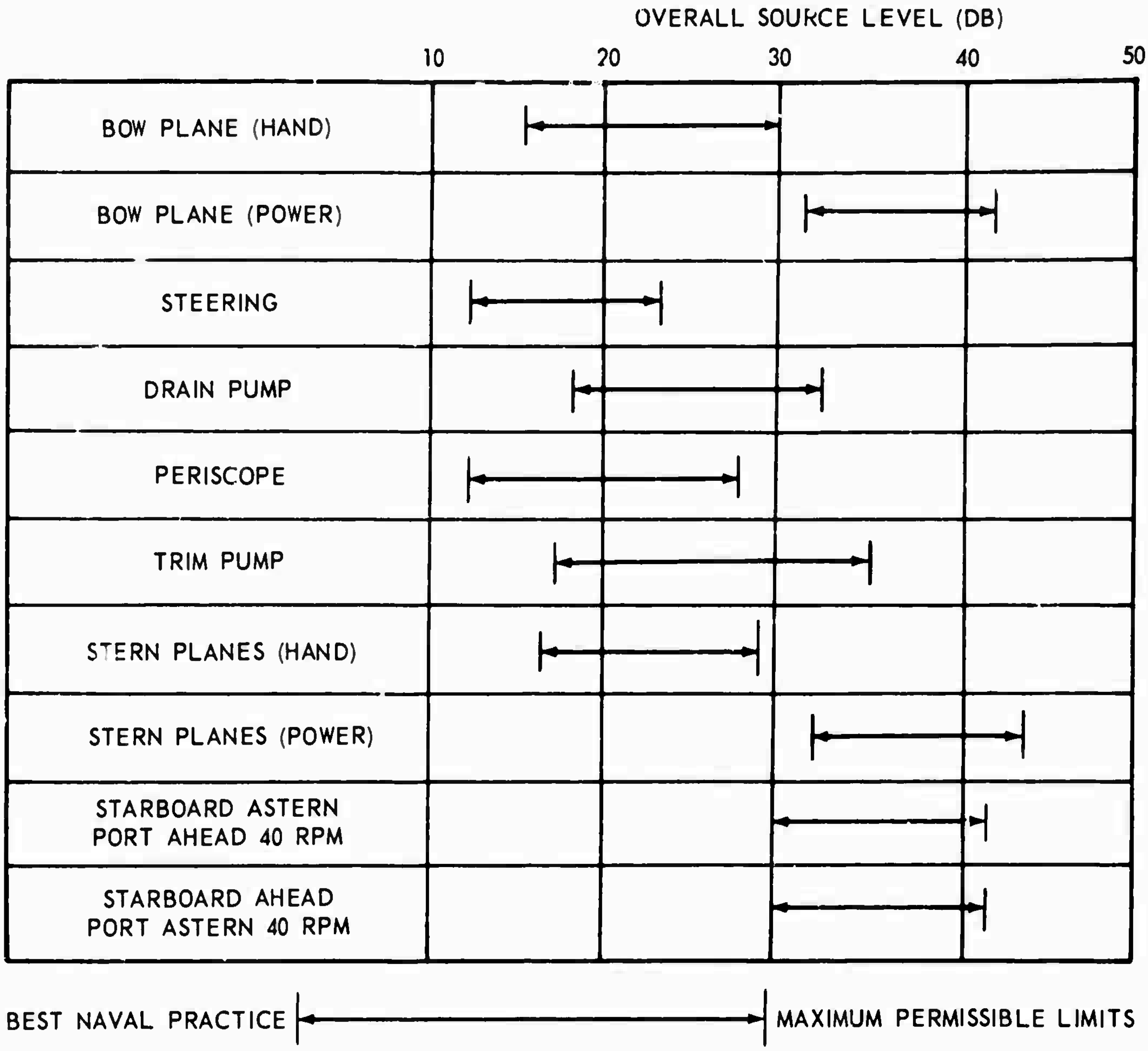


FIGURE 68. NOISE LEVEL SOURCES IN A SUBMARINE

Propeller noise is produced by cavitation at the propellers. Cavitation is the separation between the propeller blades and the surrounding water caused by the propellers turning so rapidly that the water does not close in behind the blades, thus producing a stream of bubbles. Propeller noise may be the most intense and obvious source of noise at high speeds.

Propeller noise is affected not only by speed but also by the depth of a submarine. Since the hydrostatic pressure of the water around the propeller increases with depth, a deeply submerged submarine may operate at greater speed without cavitating than one operating at a shallow depth. For the newer type submarines where the propellers are designed to give a reduction of circulation near the blades, the cavitation speed is about 40 % greater than conventional fleet-type submarines. In addition to cavitation resulting at constant forward movement of the submarine, cavitation is produced by acceleration. In summary, a particular propeller at a given depth will not cavitate unless its speed exceeds a certain critical value which is dependent on the depth of the propeller below the surface. If a propeller begins to cavitate at 50 rpm at 15 feet it will cavitate at 100 rpm at 60 feet and 200 rpm at 240 feet. Cavitation speeds for various submarines are shown in Figure 69.

Hydrodynamic noise includes, among other sources, turbulent pressures upon the transducers from flow eddies and rattles and vibrations in the submarine plating and sonar gear. The water flow around the sound gear sometimes creates the major part of self noise. Flow noise is characterized by its critical dependence on speed. This noise increases as the fifth or sixth power of speed and is independent of the operating depth of a submarine. This latter characteristic distinguishes flow noise from cavitation noise. A clean streamlined dome minimizes turbulent flow and delays the onset of dome cavitation. Measurements of total self noise on a number of destroyers as a function of ship speed show the noise level to be 6 to 10 db higher for ships whose domes are fouled by marine growth. Flow noise has a continuous spectrum, peaking in the low frequency region and increasing in intensity as the speed increases (about 24 db as speed is doubled). Low frequency, long range listening from a submarine mounted hydrophone may be seriously hampered at frequencies below 1 kc when the submarine travels at speeds greater than a few knots.

It is evident from the foregoing that the speed of a submarine and the frequency at which the noise is observed are the two most important determinants of the relative importance of self noise. The dominant areas of the various noise sources on a submarine as a function of these two factors are shown in Figure 70. At frequencies around 100 cps, machinery and flow noise are the principal sources; however, as speed increases flow noise predominates. Above 10 kc, ambient noise dominates all self noise at low speeds, and propeller noise dominates at higher speeds. At the speed where cavitation inception occurs, ambient noise is overshadowed by propeller noise. In the intermediate frequency range (1 to 10 kc) there are many combinations of speed and frequency where two or more sources may be important.

There are many sources of background noise, and those inherent in the sea itself have been collectively designated ambient noise as opposed to self noise which is unique to each ship and sound source. Ambient noise is the most intense at the lower frequencies. The energy level of ambient noise is the overall noise energy from all environmental sources. This is expressed in terms of the pressure equivalent of energy in a 1-cycle band in decibels relative to 1 dyne/cm^2. Some characteristics of the definitely known ambient noise sources are given in the following table:

Cavitation causes a rise in the self noise curve greater than 2 db per knot of speed.

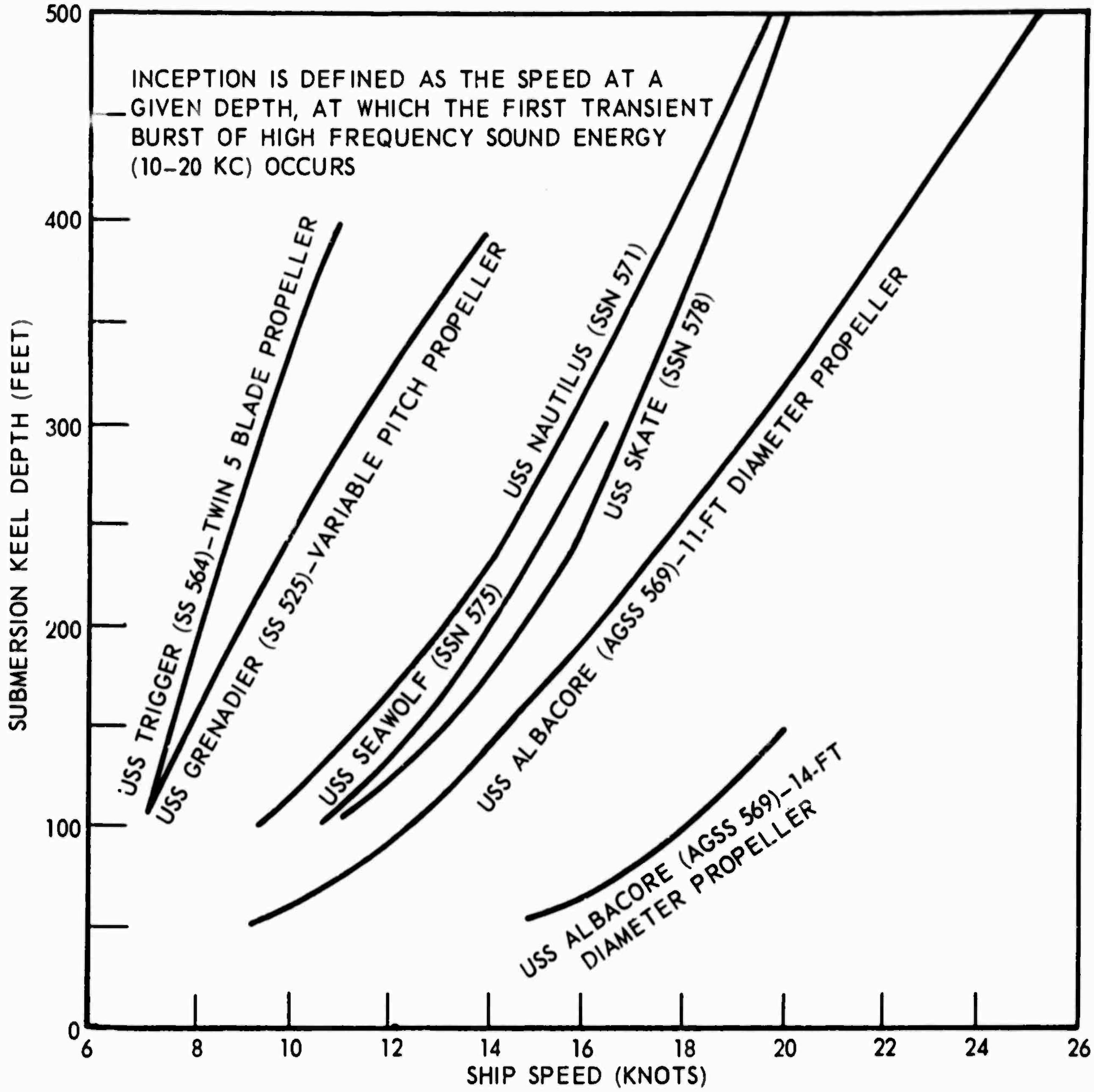


FIGURE 69. INCEPTION OF PROPELLER CAVITATION (SUBMARINES)

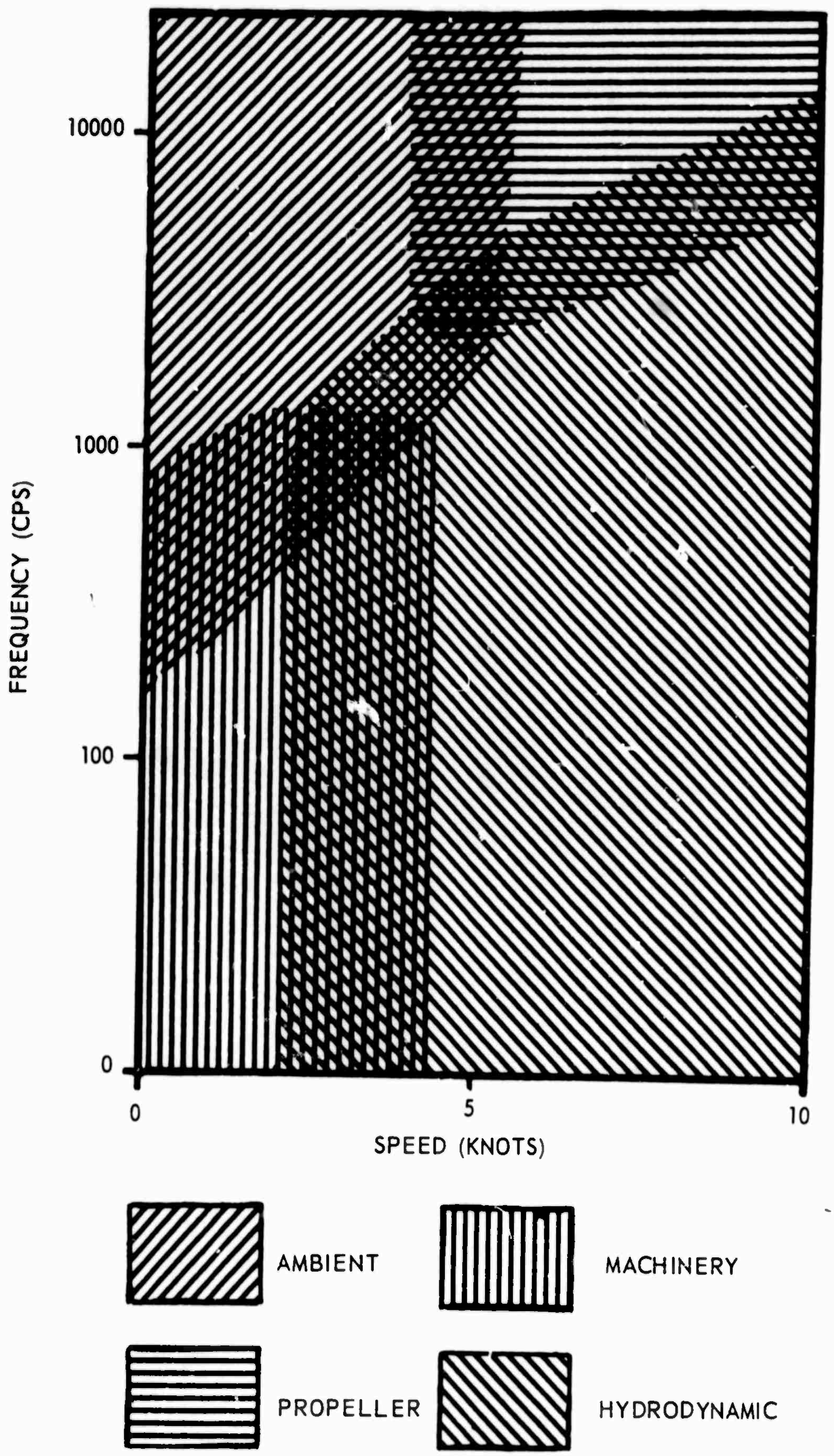
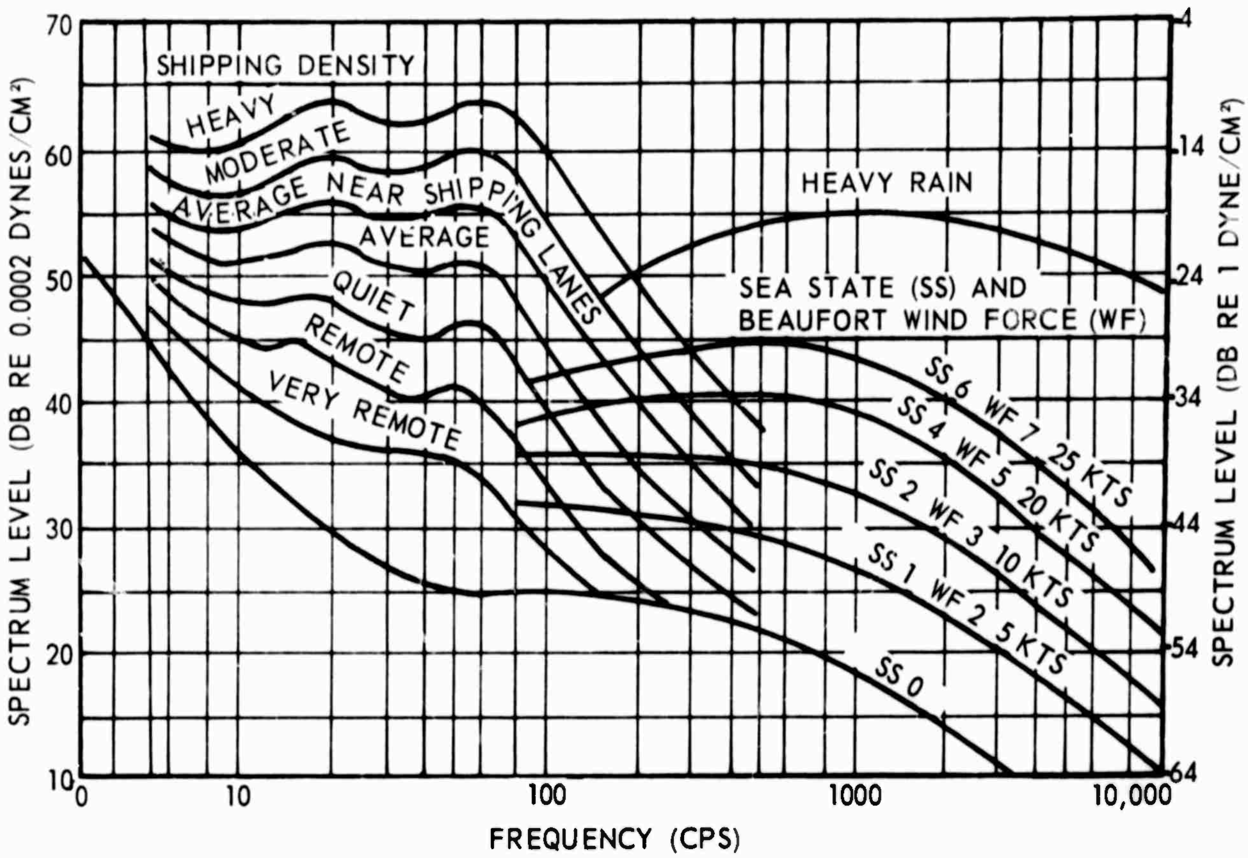


FIGURE 70. RELATIONSHIP OF BACKGROUND NOISE TO SUBMARINE SPEED



(AFTER ARTHUR D. LITTLE, INC. 1962)

FIGURE 71. IDEALIZED AVERAGE SPECTRA OF AMBIENT NOISE FOR DEEP WATER

Varying the hydrophone depth has little effect on the ambient noise spectrum between 10 and 1,000 cps as shown in Figure 72. However, at frequencies below 10 cps a decided increase in noise level occurs at a shallow located hydrophone. Since these data were taken at sea states 1 and 2, it is conceivable that the hydrostatic pressure of waves may be the contributing factor.

Sea ice can affect ambient noise levels materially in polar regions. Its influence on noise levels is dependent mostly upon the state of the ice; that is, when ice is forming, when the water surface is covered, or when ice breakup is in progress.

Provided no mechanical or thermal pressure is being exerted on the ice, the noise level generally is relatively low during the growth of ice. According to investigations carried out in the Bering Sea, the noise level should not exceed that for sea state 2, even for winds over 35 knots. This same investigation established that the intensity of the ice noise decreases with increasing frequency during the time that the ice is growing. An exception to this period of relatively low noise level is the extremely noisy condition resulting from the deformation and temporary breakup of the ice cover.

Source	Prevalence
Thermal noise, due to molecular thermal agitation of medium	At high frequencies (50 kc ±) in deep water. Forms the limiting hydrophone threshold above 50 kc ± .
Sea Surface noise, associated with waves -------------------------	In deep water between 100 cps and 50 kc ±. It is the dominant source of ambient noise in this frequency range in areas remote from coasts. Varies with sea state.
Biological noise, due to snapping shrimp and various soniferous fish and marine animals	Locally in shallow water when snapping shrimp are present; more widespread over deep water when whales and porpoises are present. Many soniferous forms of marine life are known. Most important are snapping shrimp in warm waters over rocky, coral, or shell bottoms; croakers; and porpoises.
Manmade noise, including distant ships and industrial noise in noisy harbors.	In and near harbors and shipping. Especially important at frequencies below 1 kc, where it is often the dominant source of noise.
Rain noise ---------------------	In and near rainstorms. Probably negligible below 1 kc.
Turbulence noise, due to current flow over rocky bottoms.	Often the dominant source of noise in bottomed hydrophones and mine cases at low frequencies.
Hydrostatic pressures, produced by waves above the hydrophone.	At very low (below 10 cps) frequencies in bottomed hydrophones.
Terrestrial noise, due to earthquakes, active volcanoes, microseisms, and distant storms.	In deep water at low frequencies. Speculative; importance as a source of ambient noise and not yet known.

At frequencies of 100 cps to 50 kc the sea surface is the predominant source of deepwater ambient noise. The spectrum level of ambient noise considered typical of average levels at indicated frequencies is shown in Figure 71. The noise level decreases with increasing frequency (slope of about -5 db/octave) and increases with increasing sea state. Above 50 kc, thermal agitation of the ocean controls the ambient noise level. The intensity of thermal noise depends on such factors as frequency, temperature, density of water, and sound speed. The thermal noise level is considered the limit beyond which it is impossible to measure acoustic energy in water.

Figure 71 shows the dominance of shipping noise below about 300 cps and of sea state above 300 cps. For any given set of conditions in the middle frequencies, the curves may be blended together to give one curve for that circumstance.

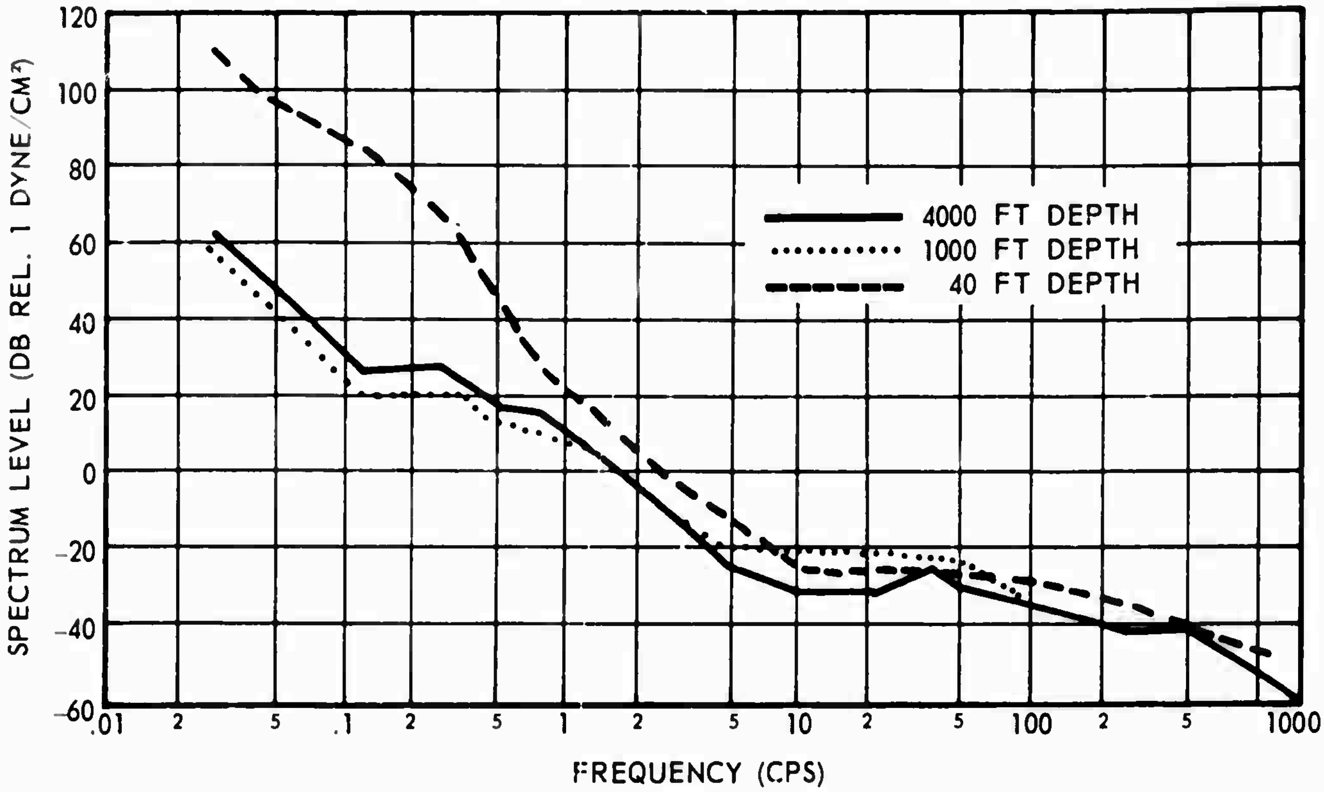


FIGURE 72. VARIATION OF AMBIENT NOISE LEVEL WITH HYDROPHONE DEPTH

The quietest conditions in ice-covered waters normally are observed during the time when the ice is neither growing, breaking up, nor being subjected to ridging and hummocking. When the formation of ice has ceased and no forces are acting on the ice, the noise level actually may be attenuated by the dampening effect of the ice cover. In the Bering Sea, for example, the average spectrum levels above 100 cps reach a minimum at this time. The ambient noise intensity at frequencies of 100 to 4,000 cps (probably to 20 kc) is comparable to a curve for sea state 1/2, even with winds in excess of 10 knots. Below 100 cps, only winds over 35 knots contribute to the spectrum. A minor contribution to the spectrum originates from the flow of water beneath the jagged underside of the ice.

Considerable noise generally is associated with the hummocking and breakup of ice. The characteristic sounds of ice undergoing stresses and strains (moaning, screeching, scraping, banging, etc.) create a high level of continuous interference to active and passive sound gear. These loud noises are comparable in level, for all wind speeds and for frequencies above 500 cps, to those existing in open water at sea state 6. They generally overshadow all other noises, except perhaps when strong winds also are blowing. Incomplete studies carried out at frequencies of 200 to 1,250 cps

indicate a positive slope of 3 to 5 db in the intensity curve between 200 and 500 cps and a negative slope of 3 to 5 db between 500 and 1,250 cps.

In general, the noise intensity in ice-covered waters reaches its maximum during the spring thaw. No correlation has yet been established between characteristics of sea ice (age, brittleness, and degree of hardness) and the equivalent sound pattern. However, sound rays traverse ice and reach the air above with greater or lesser ease according to the thickness and absorption characteristics of the ice layer. Other sources of ice noise include icebergs, ships moving through an ice field, and drift ice floating along the edge of fast ice. The following table shows sound levels (in db relative to 1 dyne/cm^2) for various conditions in the Bering Sea:

Windspeed (knots)	Frequency (cps) 20	1,250	Ice condition
2.5	-23	-53	Growth
35.0	-16	-44	Growth
2.5	-19	-31	Breakup
35.0	-12	-24	Breakup

Biological noise may contribute significantly to ambient noise in many areas of the ocean. Because of the habits, distribution, and sonic characteristics of the various sound producers, certain parts of the ocean are more intensely insonified than others. The effect of biological activity on overall noise levels is more pronounced in shallow coastal waters than in the open sea. Also, it is more pronounced in the tropic and temperate zones than in colder waters.

FLUCTUATIONS IN ENERGY LEVEL

When a sound source emits a signal that is constant in frequency and intensity and the signal is picked up by a hydrophone at some fixed distance, the consecutively received signals will differ in sound level, that is, the signal intensities fluctuate. Fluctuations arise primarily because the sound conditions along the sound path vary as the signal travels from the source to the receiver. The changes that occur in the averages of intensities of successive signals are called fluctuations. In situations (such as echo sounding) when enough time is available to evaluate echoes, fluctuation effects can be eliminated largely by averaging a successive number of pings. However, fluctuation may affect search operations when time for only one or two pings is available.

Several causes of fluctuation have been proposed, but it appears that no single cause is solely responsible for all fluctuation effects, and that all effects are present to some degree at any particular time. The following are some of the suggested causes of fluctuation: 1) Motion of the transducer because of roll and pitch of the platform, 2) focusing and defocusing effects by thermal or saline patches acting as lenses on the sound intensity over a single ray path, 3) scattering effects produced by bubbles and solid particles, 4) interference between rays traveling direct-surface; and bottom-reflected paths, and 5) interference between refracted rays produced by variations of temperature.

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APPENDIX A

CONVERSION FACTORS

1. Meters (or m/sec) and feet (or ft/sec)
   meters x 3.28083 = feet
   feet x 0.3048 = meters

2. Feet and fathoms
   feet x 0.167 = fathoms
   fathoms x 6 = feet

3. Fathoms and meters
   fathoms x 1.8288 = meters
   meters x 0.54681 = fathoms

4. Knots and meters per second
   knots x 0.51479 = m/sec
   m/sec x 1.9425 = knots

5. Knots and centimeters per second
   knots x 51.479 = cm/sec
   cm/sec x 0.019425 = knots

6. Centigrade and Fahrenheit
   $°C = 5/9\ (°F - 32)$
   $°F = 9/5\ °C + 32$

7. Miscellaneous useful relationships
   one dyne/cm^2 = one microbar
   one decibar of pressure = approximately one meter depth of water
   pressure in pounds per square inch (psi) x 2 = approximate water depth in feet

   pressure in pounds per square inch (psi) x 0.0702 = kg/cm^2
   kg/cm^2 x 14.233 = psi
   one meter depth of water = approximately 0.1 kg/cm^2
   one nautical mile = 2.027 kiloyards

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3. REPORT TITLE
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4. DESCRIPTIVE NOTES (Type of report and inclusive dates) This report is designed for use by sonar class instructors, sonar operators, and USW officers.

5. AUTHOR(S) (Last name, first name, initial)
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13. ABSTRACT
By broadening the base of fundamental knowledge about the oceans, especially underwater sound, one can be assured of improvements in the art as well as a sharing of ideas and active participation in solving problems of mutual interest. This publication provides in one source all the essential elements toward a better understanding of the medium in which the military user operates. Insofar as possible the material is presented in a simplified, easily understandable format with a generous supply of pertinent illustrations to back up the text. The use of mathematics has been kept to a minimum. The material contained herein was gleaned from many existing sources (published and unpublished), edited, and adapted for ASW use in deep water (beyond the 100-fathom curve).

122

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11 August 1966

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To: Distribution List

Subj: Errata Sheet for SP-79

1. Recipients of NAVOCEANO publication SP-79 "Oceanography for Long Range Sonar Systems, Part I, Introduction to Oceanography and Physics of Underwater Sound in the Sea," are requested by copy of this letter to make the following changes in all copies of SP-79 in their possession:

a. List of Figures, page viii. Line out entries for Figures 69 and 70.

b. List of Figures, page viii. Change page number of Figure 71 from 115 to 116.

c. Page 112. Line out last sentence of first paragraph of the page.

d. Page 112. Line out second sentence of third paragraph on the page.

e. Pages 113 and 114. Remove this sheet from the publication and destroy by burning.

f. Change page 115 to 116, and page 116 to 115. Remove the sheet with these changes and replace so page numbers read in the proper sequence.

L. E. DeCAMP
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